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SOME PETROLOGICAL AND GEOLOGICAL ASPECTS OF THE IGNIMBRITE PROBLEM [1]

by

Ye. K. Ustiyev

ntil recently the problem of "welded tuffs" ignimbrites apparently was significant only he classification of rocks and the mechan- or eruption. However, numerous discover- of the past two decades have changed the ure fundamentally. This problem has be- e one of the important areas of modern ology and has proved to be intimately rela- o the evolution of our knowledge of mag- c geology and geotectonics. In the first e, the detailed geological surveys of the war years demonstrated the very wide- ad occurrence of this still rather puzzling p of volcanic rocks. Moreover, the r role it plays in the composition of ous complexes of all epochs, from the iean to the Recent, was discovered. rally, all this attracted particular at- on and served as the basis for further overies.

oday, many features of the genesis and re of tuffs and ignimbrites are being ex- ied, but the number of obscure points and mas awaiting solution is still rather large.

IGNIMBRITES AND WELDED TUFFS

he list of unsolved problems begins with lefinition of basic concepts. The terms mbrite" and "welded tuff" themselves re- e more precise definition, for confusion een them removes one of the important s of the volcanic process from the purview trographers, as shall be demonstrated w.

s we know, the term "welded tuff" was in- uced by G. V. Abikh as long ago as 1882 he taxitic volcanic rocks of Armenia [1]. s assumed that welded tuff like the an piperno, is an acid fluidal effusive rich agmental ejecta of pumice and slag en- ped during eruptions. Similar views were held by F.Yu. Levinson-Lessing [14], P.I. Lebedev [11], and many other petrographers who studied the classic welded tuff fields on the slopes of volcanic Mt. Aragats (Alagez) in Armenia.

The situation changed after the appearances of studies by P. Marshall [45] on the acid volcanic rocks of New Zealand. Marshall demonstrated the tuff rather than lava origin of the vast sheets of sintered pyroclastic formations ("ignimbrites") in the area between Taupo and Rotorua. He held their origin to involve "incandescent clouds" like those actually observed in 1902 during the eruptions of Pelée [43] and La Soufrière [32] and that they were the hypothetical source of the partially sintered volcanic mass that filled the Valley of Ten Thousand Smokes near volcanic Mt. Katmai, in 1912 [37].

After the discovery in New Zealand, similar deposits began to be found in other parts of the world. By this time, experience accumulated in the study of this type of volcanic activity made it possible to determine that the primary motive force of the "burning clouds" or "incandescent showers" was a volcanic explosion, and the reason for sintering and even for refusion of the pulverized material was the high temperature of the gases and pyroclasts.

Clear indications of the explosive origin of rocks absolutely undistinguishable in many features from the classic welded tuff prompted the belief that welded tuffs, ignimbrites, sintered tuffs or fused tuffs (the list of synonyms numbers more than twenty) should be regarded as identical, i. e., deposits originating as showers of incandescent dust related to a special form of the explosive-volcanic process. This idea sank deep roots in petrography and was reflected in the evolution of views on the genesis of the Armenian tuff lavas, with which the entire problem takes its origin. K. N. Paffengol'tz [18] and A. N. Zavaritskiy [7] published articles offering many proofs of the pyroclastic nature of welded tuff. Today this point of view is generally shared, with a few qualifications [17, 30, 31]. We know that the sintered tuffs

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ekotoriye petrologicheskiye i geologicheskiye y problemy ignimbritov, (pp. 3-15).

of Armenia are part of a very extensive zone consisting of vast sheets extending into Turkey as far as Central Anatolia [53].

Does this mean, however, that the classification of welded tuffs of effusive origin was a petrographic error on the part of G. V. Abikh and all his successors, and that the time has come to reject this concept, limiting the nomenclature of this group of volcanic rocks to the terms "ignimbrites" or "sintered tuffs"?

Data presently available admits of a negative reply to this question.

In the discussion that developed, P. I. Lebedev [12], D. S. Belyankin [4], and later V. P. Petrov [19], noted a number of signs indicative more of a lava than a tuff origin of certain of the "ignimbrites" of Armenia. (This applies primarily to the so-called Artik tuffs.) Specifically, attention was drawn to the fact that they take the form of flow strata in which the basements are unmistakeably lavas; that there is complete structural identity with the common flow-banded lavas; and that there are no signs of an initially aggregate condition of the vitreous base. In addition, V. P. Petrov described a direct transition of vitreous lava to a frothy lithoidal pumice partially of agglomerate structure, on the assumption that the mechanism of frothing ("intumescence") can be significance in the formation of some welded of tuffs.

In accordance with these concepts, distinction is now made in Armenia between ignimbrites formed in the explosive stage of vulcanism and incandescent showers, and the quantitatively less common welded tuffs because of the distinctive form of effusion [30, 31].

Proofs of the effusive origin of certain welded tuffs have recently been found outside of Armenia. V. I. Vlodavets [6] described the welded tuffs of Mt. Semyanchik (Kamchatka), the geological and structural features of which can be explained only by assuming interpenetration of various fractions of the melt differentiated in the volcanic conduit. The subsequent effusion was apparently accompanied by partial frothing of this two-component lava and its mechanical mixture with rock fragments from previous eruptions.

M. A. Fravorskaya [28] has also described a number of cases of welded tuff formation along the coast of the Sea of Japan. Of particular interest is her observation, within a single outcrop, of a transition from a rhyolite neck lava to a typical welded tuff at the top of the deposit. In this instance, the similarity to the ignimbrites is emphasized by the presence in the light-colored rhyolite of lenticular inclusions ("fiamme") of andesitic composition. These are blocks of effusive country rock, fused and partly assimilated and separated. Similar lenticles found here and there fall into the category of conglomerates.

Serious proofs of the effusive origin of the welded tuffs are adduced by B. L. Rybalov [22] with reference to the Northern Tien'-Shan'. Here welded tuffs, lavas and tuffs of acidic composition are the products of a Paleozoic strato-volcano having a long history of activity. Judging by their bedding, the porous flow welded tuff, replete with fragmental material, erupted through the volcanic pipe and fissures on its slopes in the form of a true melt. In a number of cases they are connected by transitions to the lavas of necks, fissure veins, and small domes. Similar transitions also connect them to typical pyroclastic rocks.

Ideas with respect to true welded tuff have also been advanced in New Zealand, the home of the ignimbrite concept. L. Grange [39] described rocks ("ovaroites") whose origin is most readily explained if they are assumed to be of lava origin. The frothing mechanism of bisomatic banded lava moving in the conduit and along the slope of a volcano is fully adequate to explain the genesis of welded tuffs, in Grange's opinion.

The foaming of the melt in the conduit and its mixture with dense lava and rock fragments from the pipe walls are important features of the genesis of certain New Zealand welded lavas in the opinion of Beck and Robertson as well [34].

T. Matumoto, T. Ishikawa, and M. Minato [46] also write of the mixing of lava with pyroclastic material partially remelted in the mixing process, and with fragments of older lavas from deposits laid down by the Aso and Ayra volcanoes in Japan, and they call the resultant rock "fused lava".

In the United States, where vast fields of typical sintered tuffs have been discovered in the past decade, one sees, now and again, expressions of dissatisfaction with the concept of the incandescent avalanche as an explanation for the genesis of certain variants in the ignimbrite series. D. Hausen (see R. Smith [49]) holds that the sintered tuffs of Dorena, in western Oregon are actually frothed dacitic lavas. The base and surface of the flow are highly comminuted and resemble "pumice tuff". The central portion is a streaky blister lava with a flow texture. Biotite sheets face opposite to the direction of flow, while in the base and cover tuffs they are oriented in various directions.

J. Kennedy [10] explains the genesis of the vast field of ignimbrites in the Yellowstone Park area by the frothing of lava and subsequent compaction of the frothy mass with formation of a

...icture characteristic of sintered tuffs. In ...survey of "ash flows", R. Smith [49] pre... ...ed the opinion of F. Boyd (doctoral thesis, ...7) to the effect that the flow at the Canyon ...llowstone Park) at least could have had the ...y origin assumed by J. Kennedy.

...hus, virtually wherever ignimbrites or ...ered tuffs exist, there is evidence in the ...ks of individual flows of many features ...ch cannot be explained by deposition from ...andescent clouds", "avalanches", or "ash ...s". These are rocks which combine the ...litions of origin and structure characteris... ...both of lavas and tuffs and, therefore are in ...ordance with the definition of "welded tuffs" ...n by G. V. Abikh, F. Yu. Levinson-Lessing ... other petrographers of the first half of the ...ury. They are of effusive origin, but al...s contain a fragmentary component of ...ying composition and a number of other ...acteristics testifying to intensive frothing ...e melt and simultaneous intermixture of ...initial materials in the volcano conduit or ...e surface. Therefore, welded tuffs must ...egarded as one of the manifestations of the ...sive form of volcanism, whereas ignim...es (sintered or fused tuffs) are an ex...ssion of the explosive form.

...he geologic significance of welded tuffs ...rently is not very great and they are ...ainly less abundant than ignimbrites. ...ever, this does not detract from the rea... for drawing a petrographic distinction ...veen rocks whose appearance reflects dif...nt conditions of volcanicity and a special ...ectonic situation.

...ield experience indicates that welded tuffs ...ignimbrites are not always differentiable, ...icularly in volcanic fields, where they are ...bers of a single effusive-pyroclastic com... . Moreover, it is not possible always to ...rentiate ignimbrites from true lavas or ...e recrystallized tuffs. Nevertheless, as ...lear from the examples cited, in their ...cal form, welded tuffs, as effusive rocks, ...characterized by a number of features by ...h they may and should be differentiated ...ignimbrites as pyroclastic rocks.

... IGNIMBRITES AND GRANITE MAGMA

...n the overwhelming majority of known cases, ...mbrites (fused tuffs) are rocks of acid ...position. They usually fall into the rhyolitic-...itic or dacitic-granodioritic family, and ...frequently characterized by a somewhat ... alkaline content.

...ess frequently, the mechanism of in...descent avalanches will be found to have ...n operative during certain eruptions of an ...esitic nature, accompanied in this case by extrusive obelisks and a very small quantity of accompanying pyroclastic material. An example familiar to us is the eruption of Mt. Sheveluch in 1948, described by A. A. Menyaylov [16], which was accompanied by the formation of a massive andesite dome, from which small incandescent gas-and-dust avalanches periodically rolled down the mountain. The Pelée-type (or, more accurately, Katmai-type) eruption of Mt. Bezymyannyy in 1955-56 was on a larger scale, yielding about 2 km^3 of incandescent avalanche deposits. G. S. Gorshkov [38], who made a detailed study of this eruption, noted the andesitic composition of the extrusive dome and the pyroclastic, unsintered masses.

In exceptional cases, no examples of which are known in historic times, incandescent dust flows of andesitic-basaltic and even of basaltic composition occurred. A rare exception, in this category, is the "basic slag flow" of the Crater Lake (Oregon) caldera, containing a total of 53.9 to 56.9% silica, described by H. Williams [55]. It occurred in the culminating stage of the eruption of Mt. Mazama, which yielded approximately 30 km^3 of deposits originating as incandescent avalanches. The bulk of these constitute, however, an "ash flow" of the dacitic and rhyolitic-dacitic type. Only one-eighth of this volume comprises basalt scoria, with which the explosion ended.

The andesitic and basaltic pyroclastic deposits of Bezymyannyy and Mazama are of interest as unique examples of major incandescent avalanches, the immediate sources of which were eruption of the intermediate and basic magma. However, neither in volume nor even less in area are they at all comparable to ignimbrites of the acid, dacitic-rhyolitic composition. Volcanic fields of acid ignimbrites have now been found in all parts of the world and among continental strata of all geological ages, although, of course, those best studied are still those originating in the most recent stages of geologic history.

The thickness of such a deposit may vary from a few to hundreds of meters, and the area covered from tens and hundreds to some tens of thousands of square kilometers, and their volume from a fraction of one to tens of thousands of cubic kilometers.

The table that follows, the basic data of which are derived from materials from the latest review by R. Smith [49] and articles by J. Westerveld [51], S. Aramaki [33], T. Matumoto [46], A. I. Mesropyan [17], and I. M. Speranskaya [23], presents data descriptive of the major dimensions of certain of the best-studied ignimbritic fields. These data illustrate well the fact that a definite relationship exists between the volume of ignimbritic deposits, the type of eruption, and the geotectonic conditions to which they are related.

The incandescent clouds of volcanoes in the first group accompanied eruptions of the Mt. Pelée type and came in conjunction with very viscous magma extruded from the focus in the form of explosions that periodically destroyed the extrusive dome. Despite the very high temperature of the clouds, the ash suspended therein sintered very little or not at all, and this is expressed in the small volume of extruded matter and its rapid cooling.

Volcanoes of the second group differ from the first only in that their eruptions are of the crater form (extrusive domes do not form in all cases), and are of considerably larger scale and duration. In accordance with these features, slightly sintered tuffs are found among the deposits of incandescent avalanches. The total volume of the erupted pyroclastic material may amount to two (the Mt. Bezymyannyy eruption in 1955-56) or even five (the eruption of the Novarupta crater of Mt. Katmai in 1912) cubic kilometers. In all cases, there is an indubitable connection between acid pyroclastic material and the differentiation of the more basic magma in the focus feeding the eruption. In the majority of cases, judging by the composition of lavas of earlier eruptions, the initial magma is a pyroxene andesite in composition. The pyroclastic deposits are characterized by fumarole activity of long duration and enormous quantities of steam and gas emissions (fluorine, chlorine, volatile compounds of the heavy metals, and others).

The third category of volcanoes are distinguished by activity on an even greater scale, chiefly of the crater type, but with concomitant fissure eruptions playing a significant role. The nature of the distribution of wholly or partially sintered deposits is evidence of the existence of numerous sites of explosion, the products of which cover tens and hundreds of square kilometers and attain a volume of tens of cubic kilometers. The emptying of the magmatic chamber resulted, in virtually all known instances, in the formation of large subsidence caldera, later usually filled by lakes.

In this case, the relationship of the acid explosive magma to differentiation in the magmatic chamber is also indubitable. The pr[illegible] and subsequent eruptions (if any occurred) are characterized by lavas of andesitic and basaltic composition. In this connection, the well studied layer of partially sintered tuffs at Crater Lake (Mt. Mazama), described by H. Williams [55], is of particular interest. A large part of the stratum consists of pyroclastic deposits of acid composition, covered by a flow of "basic scoria" of basaltic and andesitic-basaltic composition (approximately 26 and 4 km^3, respectively). The clear-cut transition from the acid to the basic portion of the avalanche and the absence of clear signs of temporal separation between them indicates that there was a sharp boundary between the layers of melt differentiated in the magmatic chamber. Taken as a whole, the picture permits one to believe that a melt producing a stratum of dacite tuffs derives from the cleaving of a specifically basaltic magma.

Clear proof of the magmatic differentiation within the limits of a magmatic chamber during a precisely determinable period is provided by the eruption of Mt. Hekla (Iceland) in 1947-48. In the century since the last eruption, the melt that feeds Hekla changed in composition from basaltic to dacitic. The first explosions of the 1947 eruption yielded pumice containing 62% SiO_2, while subsequent lavas were normal for this volcano, consisting of basalt with 54% SiO_2 (8. 50).

The ignimbrite fields of the fourth group occupy thousands of square kilometers with a thickness of a few hundred meters, and a volume measured in cubic kilometers. In the geological and genetic sense, they play the role of an intermediate link between all of the preceding groups of deposits, always referable to a particular volcanic center, and the last group thereof, in which this relationship is absent or is of a substantially more complex nature.

In accordance with this, the fourth category includes both the caldera deposits of sintered tuffs in the vicinity of Mts. Ayra and Aso in Japan, and the ignimbrite fields of the Pazumakh Plateau, Sumatra, as well as the Snake River formations in the U. S. A., which relate not to specific volanic apparatuses, but to explosions of long duration and major volcano-tectonic depressions.

The last — fifth — group is doubtless that of greatest interest. A characteristic feature here is the enormous scale of accumulation of the sintered tuffs occupying thousands of square kilometers, hundreds of meters in thickness, and measuring tens of thousands of cubic kilometers in volume.

The classic ignimbrite fields of Sumatra (Lake Toba) and New Zealand (Lake Taupo) are associated with vast zones of tectonic faults and graben-like structures governing the volcanic phenomena of the central and fissure types. The elongated rift structure of Sumatra is of particular interest. J. Westerveld [51, 52] compares these in depth of occurrence and geological significance to the arc of tectonic fractures of the present-day oceanic trough surrounding the Indonesian islands arc.

A number of the Sumatran ignimbrite fields are found along this volcano-tectonic depression, among them the enormous fields of the Toba caldera and the Pazumakh volcanic highland. On its southern flank we find the

Origin, Area, and Volume of Certain Ignimbrite Deposits

Examples	Area, km^2	Thickness, m	Volume, km^3	Age	Origin
Volcanoes: Pelée (Martinique), La Soufrière (St. Vincent), Merapi (So. Java), Leamington (New Guinea), etc.	Small	Meters	0.001—0.1	Recent	Incandescent clouds, almost always associated with formation of a central extrusive cone.
Volcanoes: Komagatake, Nantai, Takahara, Asama, etc., (Japan)	Tens	Tens	0.2—2	Quaternary, Recent	Incandescent avalanches of crater eruptions, sometimes with formation of central extrusive cone.
Bezymyannyy (Kamchatka).	55—60	20—50	1.8	Recent	
Valley of Ten Thousand Smokes (Katmai, Alaska)	137	3—61	Over 4.7	Recent	
Volcanoes: Hakone, Ata, Towada, Shikotsu (Japan), Krakatau (Indonesia), Mazama (Crater Lake, Oregon, U.S.A.)	Tens-hundreds	Tens-hundreds	15—100	Quaternary, Recent	Incandescent avalanches of crater eruptions from many foci, accompanied by formation of large calderas.
Ignimbrite fields: Brisbane area, Queensland (Australia);	Over 790	61—152	47—118	Triassic	
Pazumakh Plateau (So. Sumatra).	2500	50	120	Quaternary	Incandescent avalanches associated with large caldera complexes of subsidence and fissure eruptions
Bishop tuffs, Mono Lake, Calif.	1028—1165	122—152	146	Quaternary	
Region of Mt. Ayra (Kyushu, Japan)	3870	40	154.8	Upper Pleistocene	
Region of Mt. Aso (Kyushu, Japan)	3504	50	175.2	Upper Pleistocene	
Southern Armenia	About 10,000	—	Over 100	Quaternary	
Snake River Region (SE Idaho and So. Montana, U.S.A.)	13,000	6.1—15.2	80—200	Tertiary	
Ignimbrite fields: Lake Toba Region (No. Sumatra)	25,000	Up to 600	About 2000	Quaternary	Incandescent avalanches associated with volcano-tectonic depressions and numerous sources of central and fissure eruptions, accompanied by formation of complex volcano-plutonic associations.
Lake Taupo Region (No. New Zealand).	25,900	18.3—152.5	About 8340	Quaternary	
Elkhorn Mts. (W. Montana).	7800—25,900	—	2100—4100	Upper Cretaceous	
Yellowstone Park and Absaroka Range (Wyoming, Idaho, Montana, U.S.A.).	10,360	140—285	16,680	Tertiary	
San Juan Mts. area (Colorado, No. New Mexico).	About 30,000	Up to 2100	10,000	Upper Cretaceous — Eocene	
Region of central portion of Okhotsk volcanic belt	About 25,000	—	18,000	Upper Cretaceous	
Great Basin area (So. Nevada, SW Utah, U.S.A.)	About 80,000	About 2000	About 150,000	Late Tertiary	

volcano island of Krakatau, famed for its gigantic explosive eruption of 1873 [sic].

Although they display a definite relationship to vulcanism and volcanic formations, ignimbrite fields of this type display no clear relationship to individual volcanic centers, and the conduit remains unknown in the majority of cases, as was emphasized by P. Marshall himself [45] with respect to the New Zealand ignimbrites. This is due to the very large number of sources of explosions associated with volcanic activity both of the central and the fissure type.

Despite the very large volume of acid pyroclastic material, eruptions in this case, as well, apparently represented only an episode in the evolution of volcanic foci having melts of more basic composition. Thus, the vast mass of acid ignimbrites of Sumatra predated the lengthy andesitic vulcanism, for the andesites and basalts were ejected after the ignimbrites [5, 51]. The ignimbrite field of New Zealand also lies in the Pacific zone of andesite vulcanism.

The fifth group is completed by several typical examples of the enormous Upper Cretaceous and Tertiary ignimbrite fields discovered in the past decade within the Pacific geotectonic belt. Today we know of no less than ten such — and even larger — accumulations of sintered tuffs, but all are only in the earliest stages of investigation, and no small efforts will still be required to trace their boundaries and improve the geological situation. The ignimbrite fields of this category are defined by very large zones of tectonic fractures along which extend not only the strata of volcanic rocks, but numerous granitoid intrusions of the hypabyssal type. In the majority of cases they are associated with highly upthrust volcanic highlands showing no signs of caldera formation, although instances of large graben calderas are known. Specifically, geological surveys of the transverse "porphyry belt" of the Rockies in Colorado revealed three large calderas: those of Silverton, Lake City, and Creede. According to Ratté and Steven [48], a layer of ignimbrites more than 1100 meters thick is found in the core of the caldera.

The most characteristic feature of a large number of these most enormous ignimbrite fields is the absence of signs to indicate their direct genetic dependence upon differentiation of the basic magma. The basaltic covers sometimes encountered in these areas are usually separated from the mass of acid tuffs and lavas by unconformity and considerable erosion, which may be explained by the relationship of both to various sources. The complex as a whole, including their associated intrusions, is always of acidic, rhyolitic-dacitic and granito-granodioritic composition, and is not infrequently distinguished by a high capacity to undergo mineralization. As examples we may cite the familiar cases of the iron, molybdenum, lead, zinc, silver, and gold ore manifestations in the San Juan mountain belt in Colorado [9], and the molybdenum, tin, bismuth, tellurium, mercury, and gold ores in the Okhotsk belt [25, 26].

Taken together, these features permit us to say that the enormous ignimbrite fields of this type are not genetically associated with basalt melt differentiates, the appearance of which signifies only a stage in the evolution of the primordial basic magma in the volcanic focus. There can be no doubt that they originate in true granitic (more accurately, granitoidal) magma, the derivatives of which cannot be differentiated either in composition, metallogenetic features, or geological conditions of formation from representatives of the typical "plutonic series".

Thus, these fields combine the characteristic features of both volcanic and plutonic processes. They may serve as an example of the complex "volcano-plutonic formations", the intrusive, effusive, and explosive members of which are related by uniform origin, site, and time of formation.

It is extremely interesting to note that the dimensions of the ignimbrite fields, and the scales of ignimbrite formation in general, increase rapidly as the magmatic processes change from purely volcanic to plutonic forms. This renders wholly comprehensible the fact that the largest dimensions are characteristic of fields of acid ignimbrites related to granitic magma in the narrow sense of the word.

The volumes of such fields are wholly comparable to those of the large granitic plutons. By way of example we may present the computations of A. Buddington [36] for the Potosi volcanic series in the San Juan Mountains of Colorado showing that it is equivalent in volume to a granite batholith 415 km^2 in area and 16 km thick. It will be understood that this figure does not represent an unsurpassable limit, for analogous computations for the larger ignimbrite fields would permit comparisons with granitic plutons of even larger size.

Thus, ignimbrites, or sintered tuffs, are related in their genesis to derivatives both of basic, basaltic, and acid, granitic, magmas, and it is in the latter instance that they yield fields of particularly large dimensions. As a consequence, the mechanism of ignimbrite-forming incandescent avalanches can function both in the volcanic processes proper and in the more complex ones in which phenomena of a volcanic and plutonic character are combined.

Ye. K. USTIYEV

IGNIMBRITES AND PLUTONISM

)ne of the significant geological achieve-
its of recent years is the establishment of genetic relationship between the plutonic volcanic forms of the magmatic process. s relationship finds ever more abundant ofs in the study of complicated volcano-onic complexes rising to the surface and hallow crustal depths. As it relates to the ozoides of the northeastern portion of the fic belt this problem was examined in a ent work by A. Buddington [36], and another 'e. K. Ustiyev [25] in the northwestern ion.

'he geological aspect of the ignimbrite olem is directly related to the evolution of cepts with respect to volcano-plutonic for-ions.

n the first place, it is important to hasize once again the universal spatial tionship between the largest ignimbrite ls and the associated acid lavas and tuffs, hypabyssal granitoid intrusions. In the et Union this relationship was first ob-'ed in 1949 by Ye. K. Ustiyev [24, 25], respect to the enormous Okhotsk tectonic-matic belt. The most recent detailed in-igations performed here by I. M. Speran-'a [23] have uncovered new proofs of this tionship and determined that the scale of nbrite formation was very much greater previously assumed. The formation of ck Upper Cretaceous volcanic deposit s, ignimbrites, lavas, tuff lavas) of litic composition usually preceds the em-ement of granitoid intrusions (quartz zonites, granites, alkaline granites, kites) with a number of signs indicating they are of a common origin. In some s direct transitions are observed between isive and volcanic facies.

ne can hardly doubt that large ignimbrite s will be found in an analogous geological ition in the southern portion of the East n volcanic belt, the northern half of which prises the Okhotsk member, while the hern embraces the eastern slopes of the ote-Alin' and extends across the Korean nsula and along the entire Chinese coast . The striking similarity in the develop-: of geological and magmatic phenomena in rea of the Cretaceous and Tertiary vol--plutonic complexes of the Okhotsk [24, Sikhote-Alin' [27], and Eastern Chinese zones provides a dependable basis for hypothesis.

he Cenozoic volcanic belt of island arcs contains proofs of the connection between nism and the formation of ignimbrite a. R. V. Van Bemmelen [5], who studied gnimbrite fields of Sumatra, writes of the intra-Miocene and Plio-Pleistocene eruptions of the Katmai type, which laid down thick strata of acid pyroclastic material simultaneously with the emplacement of granodioritic batholiths. The latter "almost reached the surface when emplaced."

Examples of volcano-plutonic associations with large ignimbrite fields are numerous on the other side of the Pacific Ocean as well. One of the very largest accumulations of Upper Cretaceous sintered tuffs, which lies in the Elkhorn and Boulder Mountains of western Montana, recently studied by M. Klepper and H. Smedes [42], relates to the volcanic highland belonging to the vast "transverse belt of igneous rocks" of Central Montana. The variegated Upper Cretaceous volcanic series, containing a predominance of pyroclastic deposits of quartzitic-latitic and rhyolitic composition is penetrated by the hypabyssal Laramin batholith of Boulder, consisting of granodiorites, granites, and quartz porphyries. The geological connection of the batholith with the acid volcanic rocks of this field has long since been described by B. Butler [3].

The large field of volcanic rocks in the Yellowstone Mountains, with their recently discovered, thick ignimbrite deposits [49] is also accompanied by hypabyssal post-volcanic instrusions of granodiorites and monzonite porphyries, which V. Lindgren [15] ascribes to the same source as the volcanic rocks.

The well-investigated volcanic belt of the San Juan Mountains and the Front Range of Colorado constitute one of the more remarkable instances of association of volcanic and plutonic activity, starting with the end of the Upper Cretaceous, but achieving particular development in the Eocene. Effusives and numerous small, and in some cases batholithic (Princeton) intrusions of granitoidal, and primarily quartzitic-monzonitic composition, are here associated with immense series of ignimbrites [48, 49]. The most recent intrusions display a direct relationship to explosive breccias [36], and consequently were formed virtually at the surface.

In all such cases, the relationship between ignimbrites and plutonic action is confirmed not only by structural, spatial, and chronological association, but by a number of features testifying to a common petrographic nature. Moreover, in ignimbritic series one sometimes finds fragments of intrusive rocks of exactly the same composition as the intrusive ignimbrites. H. Williams [54] has described the welded deposits of rhyolitic ash showers from the Navaho country in northeastern Arizona, which are full of granite fragments. G. McCall [44] encountered rounded fragments of acmitic syenite in the trachyitic ignimbrite of the Menengai caldera in Kenya. Identical intrusive syenites

are known to exist in Mt. Kenya and Mt. Mawenzi of the Kilimanjaro massif. The author assumes a syenite intrusion to exist beneath the Menengai caldera as well. Rounded fragments of intrusive rocks were encountered by J. Westerveld [53] in the ignimbrites of the Konya and Beysehir areas of Turkey. The composition of the ignimbrites corresponds to that of the Yosemite granite; the composition of the fragments is unfortunately unknown. The relation between the ignimbrites and plutonic action has yet another form, which cannot be left out of consideration.

In recent times, complicated vulcano-plutonic complexes are found ever more frequently. In these, central intrusions and ring and conical dikes surrounding them are associated with tuffs and lavas. A splendid example of such a ring structure has recently been described in detail by K. N. Rudich [21], in the central portion of the Sarychev range in the Yana-Kolyma folded region. A large dacitic subvolcano is encircled by ring dikes of granodiorite porphyries and an outer belt of tin-bearing granitoidal intrusions. Within this enormous ring structure (200 x 300 km) there have been found large fields of rhyolites, dacites, and the corresponding tuffs.

A highly interesting example of combination of an ignimbrite deposit with a ring intrusion was recently studied in the Kuraminskiy Range by V. A. Arapov and V. N. Tkachev [2 and oral communication]. In this instance, a large isometric field of dacitic ignimbrites is bounded by a semicircular intrusion of granodiorite porphyries of the identical petrographic and petrochemical type.

It is necessary to point out that in a number of cases, the "tuffs" and "lavas" of such ring complexes are actually ignimbrites. Specifically, H. Williams and J. Thompson (see A Buddington [36] and R. Smith [49]) proved, not long ago, that the renowned complex "lopolith" of Sudbury in Canada is actually a volcano-tectonic depression enclosed by a ring complex of dikes, sills, and eruptive centers. The depression is filled with an enormous series of ignimbrites, previously held to be rhyolites and tuffs. The volcanic complex covers the granophyric granites and granodiorites of the upper portion of the "lopolith" with which they are apparently connected in places by gradual transitions. The total volume of the ignimbrite strata here attains 750 km^3 in an area of about 480 km^2.

The deposits of the Permian volcanic plateau in the vicinity of Oslo, Norway, previously believed to be alkaline trachytes (rhomb-porphyries) and rhyolites have also proved to be ignimbrites. According to the latest data of C. Oftedahl [47], the formation of ignimbrites in this case, as well, is associated with major collapse calderas, intrusions of larvikites, syenites, and granites, and ring dikes of quartz porphyries.

M. Billings [35], generalizing data on ring structures and the intrusions related thereto, drew attention to the fact that 11 of the 30 examples described to that day are characterized by a combination of plutonic and volcanic rocks. It now appears that certain of these ring complexes reveal the presence not only of lavas but of thick ignimbrite series. This is evidence that the formation of ring and central intrusions was accompanied not only by effusive but by an explosive form of the magmatic process, followed by tectonic subsidence and the formation of the caldera. An important element in this chain of events is apparently the emplacement "almost to the surface itself" of a central granitoidal pluton, which culminated in an explosion which caused the cover of the intrusion to be destroyed and the accumulation of incandescent pyroclasts, and ignimbrites, whose volume was larger as the volume of the magmatic body increased and, consequently, the more powerful the explosion.

The coincidence of these processes and their final result point to a new approach to the study of ring structures and, specifically, to the need to seek signs of the powerful explosions involved and of ignimbrite formation. The ignimbrites themselves are of special interest in the light of these data. Having appeared at the surface as a consequence both of volcanic and of plutonic processes, they perform the function of a connecting link between these forms of magmatic phenomena.

F. Yu. Levinson-Lessing [13] coined the phrase "unsuccessful volcanoes" (i. e., volcanoes not reaching the surface) for certain near-surface laccoliths of the Caucasus. By the same token, many of the ignimbritic series referred to above might be termed granitoidal plutons that have broken through to the surface.

CONCLUDING REMARKS

1. Incandescent ash avalanches and the ignimbrites resulting therefrom are related both to the volcanic and the plutonic forms of magmatism. The greatest dimensions are attained by the ignimbrite fields included in the complex volcano-plutonic formations of the Okhotsk-type volcanic belt in the Soviet Union or the San Juan Mountains — Front Range volcanic belt in the U. S. A.

2. Ignimbrites due to volcanic phenomena in the narrow sense of this term are, in the majority of cases, clearly dependent upon the differentiation of magma of basic or intermediate composition. This relationship is confirmed both by the geological situation in the area wher

mbrite deposits of this type occur, and the ely observed, direct transitions between mbrites of acid and basic composition.

gnimbrites associated with volcano-plutonic nplexes are just as definitely dependent upon nitoid magma, the derivatives of which d enormous masses of acid rocks in in-sive, effusive, and explosive facies. The umes occupied by these complicated com-xes yield nothing to those of certain classic anitic batholiths." Identically, there are no erences between the derivatives of this gma and the "typical granitic magma" with pect to their metallogenetic features and ability of mineralization.

3. Taken together, all this testifies to an ubitable genetic relationship between the canic and plutonic forms of magmatism, ticularly if this problem is regarded in ns of the geological distribution of the upper onic levels of the earth's crust.

Thus, the conceptions of W. Kennedy and E. erson [41] with respect to two independent ociations of igneous rocks — volcanic and onic — is clearly in contradiction to the logical facts. The idea advanced by them supported by G. Reed [20] and certain er geologists with regard to the need to arate igneous rocks into two separate sses does not reflect the geological rela-ships between volcanic and plutonic associ-ns and, therefore, is unacceptable.

4. One is led to believe that the discoveries ecent years with regard to the distribution genesis of ignimbrites will increase the rest therein of a large section of the Soviet logical profession and will facilitate the overy of large new fields in the area of young and old folded zones of the U.S.S.R. ust be noted that the ignimbrite series in plicated volcano-plutonic complexes are only of scientific but of practical interest, icularly in the light of their possible ntial relationship to shows of ore geneti-y related to granitic magma.

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RESULTS OF SEISMIC-SONIC INVESTIGATIONS IN THE AREA OF THE DEEP-WATER JAVA TRENCH[1]

by

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The seismic-sonic investigations in the area of the Java Trench made during the 31st voyage of the research ship Vityaz', were carried out by the reflection method using a multi-channel receiver.

The deep-water Java Trench runs parallel to the Greater and Lesser Sunda Islands for a distance of about 1800 km. Its maximum width is about 40 km, including only the abyssal portion below the 6500 m isobath, and its maximum depth is 7450 m. The Java Trench is separated from the Greater and Lesser Sunda Islands by the Bali Trough and the outer submerged ridge of the Javan island arc with a minimum depth of 3500 m above it. The submerged slope of the Java island arc is characterized by considerable steepness on the ocean side, amounting to 17.5°, and is of rather uniform relief.

The seismic-sonic profile of the Java Trench began at a point south of the western portion of the island of Bali (Lesser Sunda Islands) on the submerged slope of Java (Figure 1), then intersected the Bali Trough, lying between the outer and inner ridges of the Javan arc. The maximum depth of the Bali Trough in the area of the seismic-sonic studies was 4.5 km, according to the echo-sounder. Then the profile passed through the Java Trench, 7200 meters deep in this area, intersecting its outer wall, and terminated in a trough in the ocean floor approximately 650 km from Bali.

On the submarine slope of the Island of Java, the Bali Trough region, and that of the Java Trench, seismic-sonic stations were spaced at intervals of 18.4 km, while past the trough the interval was 75 km. The total length of the profile was 475 km.

As we know, seismic studies were made in the Indian Ocean during the round-the-world voyage of the British survey ship Challenger in 1950-1952 [16]. The method and apparatus developed by M. Hill [19] were employed in these studies in which measurements were made by refracted waves using radioacoustic buoys. The maximum length of a profile was 37 km and was governed by the range of the transmitters on the buoys. The explosion was set off at a depth of 270 m. After a profile had been made the vessel doubled back and picked up the buoys. It was only very rarely that profiles of this length made it possible to detect the bottom limit of the earth's crust, characterized by a velocity of propagation of the longitudinal wave of 8.0 km/sec.

The principal investigations on this expedition were conducted in the Pacific Ocean. Only five stations were observed in the Indian Ocean these being in the central portion south of the equator. Rocks with seismic velocities of 6.3 to 6.7 km/sec were found at virtually all ocean stations. They were covered with sediments 0.5 km thick [16].

The method of reflected waves using a multichannel receiver system permits study of the sedimentary series in relative detail. The refraction shooting method is employed chiefly for abyssal seismic sounding of the earth's crust.

The method of investigation we employed is generally analogous to that previously used by the Institute of Terrestrial Physics of the U. S. S. R. Academy of Sciences [2] in its studies during the Pacific Expedition of 1958 and the work of the Institute of Oceanology of the U. S. S. R. Academy of Sciences in the 30th voyage of the Vityaz' in 1959 [5]. The distances between seismic stations in the studies performed during the 31st voyage of the Vityaz did not permit the running of a continuous system of observations as was the case of the seismic petroleum prospecting at sea [15], and therefore only tentative boundaries of the levels reflected were entered when sections were plotted. Therefore, sonic oscillations on a multi-channel receiver system were alternat[ed] with reception on two hydrophones, as was don[e]

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[1] Rezul'taty seysmo-akusticheskikh issledovaniy v rayone Yavanskogo Glubokovodnogo Zheloba, (pp. 16-25).

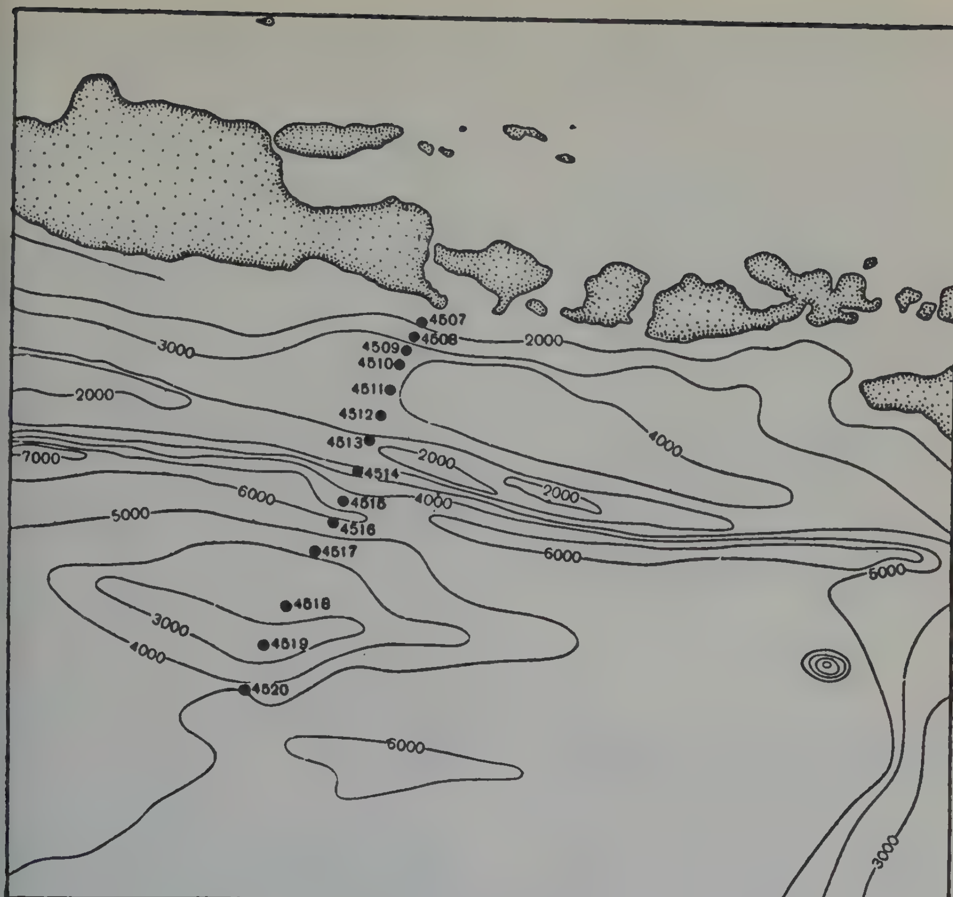


FIGURE 1. Location of Stations Across the Java Trough

e seismic investigations performed by the tute of Oceanology of the U. S. S. R. demy of Sciences abroad the Vityaz' in [9].

he frequency-response curve of the first iver sections of the multi-channel system, , the matching transformer — pressure iver made it possible to receive audio llations in the 5 to 200 cycle range [12].

he acoustic oscillations received by the ophones were amplified by the U-6 lifier of the SS-24-P seismic station, em- ng frequency filtering, and was recorded POB-14 magnetic oscilloscope with the ographic paper moving at 12 cm/sec. The ption of sonic waves by the multiple- nel receiver was based on filtering of the 30 cycle waves. This made for the best ification of the first incoming reflected s from the reflecting levels in the bottom- beds. A number of different types of ring in the 30 to 200 cycle range were set up to receive the sound waves on two hydrophones (Figure 3). The recording of elastic oscillations with various filtering arrangements will make it possible, as the data are subjected to further elaboration, to determine the relationship of the coefficient of reflection to the frequency in the vertical incidence of the sound beam.

The elastic waves were produced by explosions of charges of TNT of from 2 to 10 kg. The depths at which the charges were set off were selected on the basis of computation of elimination of pulsation of the gas bubble and amounted to between 1. 2 and 2. 4 meters [4, 10]. The TNT charges were set off by electrical detonators using the BM-52 blasting machine, with electrical recording of the moment of blast on the oscilloscope. Recording of time was by signals provided by the MKh-6 contact chronometer and a time recorder in the POB-14 oscilloscope.

After the explosion, the hydrophones receive

not only direct "aqueous" waves and waves singly reflected from the ocean bottom and subfloor levels, but multiple reflected waves. The observations always recorded second reflections, which were then employed to determine coefficients of reflection. Inasmuch as, with multiple reflections, the major losses of intensity occur upon reflection from the ocean bottom, it was possible, in our researches, to ignore losses from the water-air interface, and we took the coefficient of reflection in this case to be unity. It may be assumed that the errors are very insignificant, inasmuch as the coefficient of reflection from the surface of the water is 0.9994 [3].

The characteristic of the reflecting properties of the ocean floor and the subfloor levels can be derived by comparing the amplitudes of singly reflected waves to the amplitudes of waves twice reflected from the bottom (subfloor levels) and once from the surface of the water. Thus, we obtain a formula for the coefficient of reflection for vertical incidence of the beam of sound:

$$K = \frac{2A_2}{A_1}, \qquad (1)$$

where A_1 and A_2 are the amplitudes of the first and second reflections from a single level. Using Rayleigh's formula, we obtain:

$$\frac{2A_2}{A_1} = \frac{\rho_2 C_2 - \rho_1 C_1}{\rho_2 C_2 + \rho_1 C_1}, \qquad (2)$$

where ρ_1 and ρ_2 are the densities of the rocks on both sides of the interface, and C_1 and C_2 are the speed of sound in these strata.

This formula demonstrates that the intensity of the reflected wave increases with an increase in the difference in acoustic hardness. The density varies within considerably smaller limits than the velocity of the elastic waves, and therefore the coefficient of reflection is determined primarily by the velocity ratio of the rocks on both sides of the interface [3].

If the elastic properties of the medium do not undergo a sudden change, there will be a transition layer in which acoustic hardness changes smoothly from the value $\rho_1 C_1$ to $\rho_2 C_2$. Therefore equation (2) is valid only for waves whose length is long relative to the thickness of the transition layer, whereas for short waves the coefficient of reflection approximates zero, since this medium can be regarded as continuous with respect to short waves [3]. From this it follows that, in some cases, it is advantageous to record reflected waves by means of equipment sensitive to the low frequencies. These conclusions have been confirmed in practice in seismic studies at sea. The best pass band was the interval from 20 to 50 cycles. When the work was done at higher frequencies, in the 200 to 500 range, the recording of the reflected waves was indeterminate.

The waves in the 0 to 30 cycle range were filtered out in the reception of sonic vibrations on the multiple-channel system. The values of the coefficients of reflection were determined independently in terms of the wave amplitude received by each hydrophone in the receiver system, whereupon the mean values of the coefficient of reflection for each point at which sound waves were received, were calculated. For purposes of further interpretation of the materials obtained by these studies, and for characterization of the reflecting properties of the ocean floor and the subfloor levels, the values of the coefficients of reflection were averaged for each interface over the entire length of the profile. The computed mean coefficients of reflection are presented in Table 2.

It must be pointed out that the values of the coefficients of reflection obtained are not the true, but the effective values, inasmuch as they characterize the reflective properties of the medium with respect to waves whose length is comparable to the thickness of the strata studied.

We used the values obtained for coefficients of reflection not only to characterize the reflective properties of the surfaces, but for approximate computation of the velocities of the elastic waves in the ocean-floor strata studied. The estimates of sound velocity for the deeper layers are less exact, inasmuch as the values of the densities of the rocks of which these levels are composed, are unknown. Comparison of data due to seismic investigations and on gravity anomalies led Wurzel and Schubert to conclude that a standard column of the continental crust consists of rocks 33 km in thickness having a mean density of 2.84 g/cm^3, overlying a shell the mean density of which is 3.27 g/cm^3. A standard column of oceanic crust consists of a layer of water 5.0 km thick, of 1.03 g/cm^3 density, a sedimentary layer 1.0 km thick with a density of 2.3 g/cm^3, and crustal rock (tentatively called basalt) 4.5 km thick, with a density of 2.84 g/cm^3. The lower limit of the crust is at a depth of 10.5 km beneath sea level, below which there is a subcrustal substrate, the density of which is 3.27 g/cm^3 [21]. With the aid of these density data, as well as laboratory measurements of the density of the upper stratum of precipitates obtained by bottom dredges and continuous coring devices, and knowing the velocity of sound in the layer of water above the ocean floor and in the upper stratum of sedimentaries, as well as the mean values of the effective coefficients of reflection, we computed the velocities of sound in the ocean floor strata.

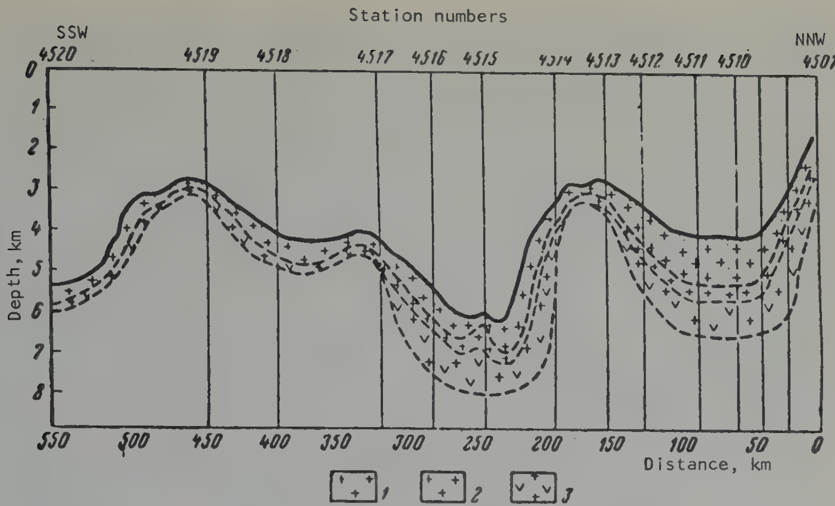


FIGURE 2. Seismic Profile Across Java Trench

1 - semi-consolidated sediments (velocity 1.7-2.5 km/sec); 2 - lithified or volcanic sediments (velocity 4.0 km/sec); 3 - igneous sediments (velocity 5.6 km/sec).

Correlation of the waves reflected from sub-oor levels was based on the shape and intensity the record, as well as on the effective coeffi-ents of reflection.

After the materials were processed, a ntative section was plotted (Figure 2). The stance between stations along the horizontal is, and the vertical depth of the reflecting rata beneath the floor were laid off in kilo-eters.

Four to six reflecting levels may be dis-nguished on the seismograms derived in the ea of the Java Trench. The seismic waves eflected from these levels could be followed virtually all stations and may be correlated th satisfactory confidence for adjacent points observation.

The laboratory determinations of the density the earth for the upper layer of the sedi-ents in specimens obtained by bottom dredges d continuous coring devices in the vicinity of e Java Trench, yield values of from 0.9 to 8 g/cm^3.

The mean density was taken to be 1.4 g/cm^3. ydrological data were employed to determine e velocity of sound in the water at the bottom, hich came to 1.53-1.55 km/sec. The velocity sound in the uppermost sedimentary layer, hich proved to be between 1.6 and 1.7 km/sec, as determined from the effective coefficients reflection, the values of the speed of sound the layer of water nearest the floor, and the ean density of the uppermost sediments.

This value for the velocity of sound was confirmed by measurements with an ultrasonic seismoscope. These measurements were made using samples of the bottom obtained at various points in the Indian Ocean by continuous coring and heavy plunger coring devices (Table 3). The mean length of a column of bottom material was from 2 to 4 meters. The maximum — 15 meters — was obtained with a heavy plunger coring device at station 4634.

The mechanical and material composition of the sediments in which measurements of the velocity of sound were made by seismoscope corresponds essentially to the composition of silts brought up in the vicinity of the Java Trench. Measurements were made by running a line of profiles in order to increase their accuracy [11].

The first measurement of velocity employed a base line of minimal length between receiver and transmitter (10-15 mm), while those which followed were made with the base gradually lengthened 15 to 20 mm at a time, the last base being 170 to 190 mm long. In all measurements, the sound wave receiver was stationary. After a series of measurements of velocity in specimens of bottom materials, curves were plotted for the relationship of sound to the length of the sample. The distance between the receiver and the transmitter were laid off along the abscissa axis, while the time required for passage of the ultrasonic oscillations, read off against the time marks in the seismoscope, was plotted on the ordinate.

Calibrating measurements of the velocity of sound in distilled water, with simultaneous temperature determination, were performed to increase the accuracy of the measurements. These seismoscope measurements of the velocity of sound were compared to its velocity in the same water, but computed on equations [7]. The correction obtained was introduced in the measurement of velocity in specimens of ground. These calibrating measurements were made before and after the measurement of the velocity of sound over the entire length of the core. The temperature of the core material was measured during the sound velocity determinations. We assumed, with some approximation, that the temperature of the sediments right on the ocean floor was equal to that of the bottom layer of water determined by hydrological observations. Thus, proceeding from the difference between the temperatures of the bottom layer of water and the temperature of the ocean floor, one may, at the time the measurement is made, introduce the appropriate correction for temperature into the measured value of the velocity of sound.

The temperature of the floor at the moment the measurements were taken is usually 25 to 27°, and the temperature of the bottom level of water is about 2°. The computed correction was usually 80 meters on the average, or 5%. If we take into consideration the accuracy of measurements by the ultrasonic seismoscope, which come to 1.0% [11], the introduction of such a correction proves to be desirable. The averaged results of the measurements of the speed of sound by seismoscope in various bottom materials are presented in Table 1.

It is of interest to observe that the values we obtained for velocity in the top layer of Indian Ocean sediments are very close to those we found in the Black Sea, and to the data for various parts of the world ocean published by L. Hamilton [17].

The seismic waves reflected from the first reflecting layer (the ocean floor) are characterized by a low intensity of recording (Figure 3b). The effective coefficients of reflection are equal to 0.10-0.20.

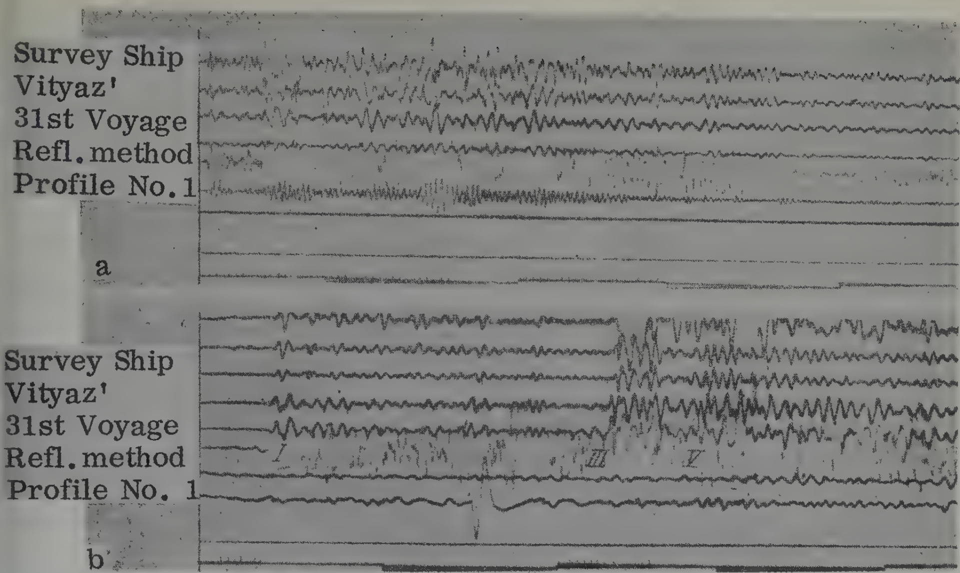
The first arrivals of seismic waves reflected from the second layer lag the first arrivals, reflected from the ocean floor, by 0.06-0.09 sec (Figure 3b) and may be followed virtually over the entire line of profiles. This layer is characterized by an effective coefficient of reflection of the order of 0.15-0.20 and by an intensity of recording approximately equal to the intensity of recording of the waves reflected by the ocean floor.

The averaged value of the velocity of sound for the upper layer of the sedimentary strata, computed on the effective coefficients of reflection, and measured by seismoscope, are taken to equal 1600 m/sec. The thickness of the surface layer of sediments was found to be 50 to 80 meters thick, as obtained from the time of arrival of the first waves reflected by the second layer, and the computed mean speed of

Table 1

Station No.	Brief Description of Sediment	Collected by:	Core length, cm / Velocity, m/sec
4623	Silty calcareous foraminiferal ooze	Continuous coring device.	395 / 1540
4626	Same	Same	309 / 1530
4630	Silty-clayey calcareous foraminiferal ooze	Same	438 / 1580
4634[1]	Same	Heavy plunger-type coring device	1503 / 1530 — 1580

[1] The heavy plunger-type coring device was used at station 4634 to take a core 1503 cm in length. The velocity column shows two values: one the mean velocity in the upper portion of the core, and the second, — the mean velocity of sound in the lower portion of the core.



a - variously filtered; b - first multiple: I - waves from first reflecting bed; III - waves from third reflecting bed; V - waves from fifth reflecting bed.

ground. It would seem that this layer consists of uncompacted bottom-set deposits. It is interesting to observe that a layer of approximately this thickness and having these characteristics is also observed in other portions of the Indian and Pacific Oceans where we made seismic investigations.

The first arrivals of seismic waves reflected by the third layer are characterized by the highest intensities recorded (Figure 3b). The effective reflection factors display elevated values, attaining 0.30-0.45. These waves correlate convincingly over the entire line of profiles. The velocities of sound computed on the effective reflection factors and assumed to hold for this layer of sediments of 1.8 g/cm^3 density, equal 2.5 km/sec on the average. These data are confirmed by studies previously performed in the ocean by refraction methods [6, 12, 14].

The second stratum appraently represents the semi-consolidated deposits of bottom-set beds. The thickness of the sediments of the second layer are not uniform along the line of profiles. On the submarine slope of the island of Java, the thickness of this layer is about 400 m, whereas it gradually increases southward toward the Bali Trough. The maximum thickness of the second layer of sediments was determined to be 1400 meters, at the bottom of the Bali Trough, while on the underwater peaks of the Java island arc it declines considerably, to approximately 300 or 400 meters.

On the slope of the Java Trench, the thickness of the second bed again increases, attaining a thickness on the order of 1000 meters at the bottom of the trench. Toward the ocean, it diminishes gradually, and, at the bottom of the depression 650 km from the island of Bali, the thickness of this layer is 300 meters.

The first arrivals of seismic waves reflected by the fourth bed lag behind the first arrivals of waves from the third bed by 0.10-0.12 sec (Figure 3).

The correlation of these waves is assuring as far as station 4512 on the south side of the Bali Trough. The effective reflection factors are 0.25-0.50. If we assume the density of this layer to be 2.2-2.6 g/cm, the mean velocity therein will be approximately 4.0 km/sec, and the thickness of the layer will be 200 to 250 m. It is possible that the third layer consists of the lithified deposits of bottom-set beds or volcanic material. This layer probably corresponds to what is termed the "second layer", first discovered in studies of the Atlantic [18]. The majority of investigators have interpreted this layer as volcanogenic. Sedimentary origin would appear to be less likely, inasmuch as velocities here proved to be excessive for this — from 2.1 to 5.5 km/sec.

Analysis of reflected waves in seismic investigations and laboratory measurement of the velocity of sound in specimens of sedimentary origin [11, 17] have revealed a noticeable increase in the velocity of sound with an increase in the depth of occurrence of the sedimentary layer, related to the compaction of the sediments. L. Hamilton's determination of the sound velocity gradient yielded 0.5-3.0 sec^{-1}. It may be hypothesized in that connection that the layer found in our investigations consists of compacted (lithified) bottom sediments.

The first reflections from the fifth bed lag behind the first reflections from the fourth by 0.18-0.20 sec. In intensity and shape of the graph, this reflecting bed correlates convincingly, until station 4512, in the southern portion of the Bali Trough, is reached. Thereafter, the correlation is less convincing because of the complexity of the graph resulting from reflection of the waves from the highly dissected surface of the ocean floor and the subfloor strata in the area of the Java Trench and the elevations adjacent thereto.

The reflection factors computed are from 0.25 to 0.50 and hardly differ from those of the overlying layer. The velocities of sound we found in the fourth layer, with density of 2.8 to 3.0 g/cm^3, do not exceed 5.6 km/sec. It is possible that the fourth layer, 1000 meters thick, is of volcanic origin.

The first arrivals of waves reflected by the sixth bed are nonuniform in intensity and nonidentical in the shape of the graph.

The value of the effective reflection factors are scattered in the interval from 0.1 to 1.0. It would appear that these characteristics resulted from the impossibility of employment of our apparatus and method in the investigation of reflecting beds at depths of more than 8 km, including the thickness of water, and also because of superposition of multiple reflections in these beds. Therefore, the characteristics derived cannot serve to provide an estimate of the reflecting capacities of the beds under study (Table 2).

The total thickness of the sediments in the area of the Javanese island arc, as we determined it, varies from 0.4 to 2.5 km. On the underwater slope of the island of Java the thickness of the sedimentary layer is 1.5 km. It increases gradually toward the Bali Trough. The maximum thickness of the bottom deposits was found to be 2.5 km on the floor of the Bali Trough. At the peaks of the mountainous elevations adjacent to the Java Trench, the thickness of the sedimentary layer drops sharply to 0.5 km, then increases again on the slopes of the trench, and attains 2000 meters at its bottom. Toward the ocean, the thickness of the sedimentary layer diminishes, attaining 0.4 km at the bottom of the depression 650 km from the island of Bali.

The climatic conditions of the mountainous islands of Indonesia favor the delivery of a great deal of terrigenous material from the land. The present high activity of the volcanoes of the Greater and Lesser Sunda islands (and particularly the catastrophic eruption of Krakatau in 1883) favor abundant formation of fragmental ejecta in the region of the Javanese island arc.

During its 31st voyage, the Vityaz' set up some thirty stations in the vicinity of the Java Trench at which sepcimens were taken from the ocean floor. On the submarine slope of the islands of Indonesia, in the vicinity of the Bali Trough and in the Java Trench, sedimentation of terrigenous origin predominates: silty-clayey and clayey oozes. On the bottom of the Java Trench, where the depth is 6 km, the clayey sediments right on the ocean floor are

Table 2

Reflecting bed	Effective reflection factor	Density g/cm^3	Velocity, km/sec	Thickness of stratum, meters	Composition of rocks of bed
I (floor)	0.10—0.20	1.03*	1.53—1.55*	—	Water at the ocean floor
		0.9*—1.8 (1.4)	1.6—1.7	50—80	Uncompacted sediments
II	0.15—0.20	1.8	2.5	800—1400	Semi-compacted sediments
III	0.30—0.45	2.2—2.6	4.0	200—250	Lithified or volcanogenic sediments
IV	0.25—0.50	2.8—3.0	More than 5.6	1000	Volcanogenic sediments
V	0.25—0.50	—	—	—	—
VI	0.10—1.0	—	—	—	—

*Measured values

Table 3

Coordinates of Stations at Which the Ultrasonic Seismoscope Was Employed

Station No.	Start of station		End of station	
	Latitude	Longitude	Latitude	Longitude
4603	16°05'0 S	76°16'1 E	16°04'8 S	75°57'0 E
4619	9°17'2 N	75°57'0 E	9°11'3 N	75°53'5 E
4620	8°34'2 N	75°38'4 E	8°26'9 N	75°39'8 E
4623	4°49'4 N	74°12'1 E	4°43'7 N	74°03'4 E
4626	4°26'0 N	72°33'1 E	4°22'6 N	72°24'9 E
4630	3°11'0 N	67°02'1 E	3°10'3 N	67°02'3 E
4632	0°09'5 N	66°44'0 E	0°06'9 N	66°36'0 E
4634	2°46'8 S	65°41'8 E	2°49'0 S	65°44'5 E
4649	16°03'1 S	62°49'8 E	16°03'4 S	62°43'1 E
4653	16°04'6 S	57°54'5 E	16°02'4 S	57°45'0 E
4660	14°59'8 S	53°38'4 E	14°54'1 S	53°35'1 E

ually noncarbonate. In the Bali Trough, of 4 km depth, they are very slightly car- ate. Calcareous sediments — foraminiferal es — are very pronounced at these depths in r parts of the ocean. In the vicinity of the nd of Java, sediments of volcanic origin in form of pumice and volcanic scoria have been found. The rapid rate at which sedi- ts accumulate in the region of the Javanese nd arc is demonstrated by the high boundary the oxidation-reduction interface in the sedi- ts. Beyond the limits of the trough, the kness of the oxidizing layer increases, and osits of terrigenous origin are supplanted alcareous and foraminiferal oozes [1].

he results of the seismic-sonic investiga- s demonstrated the accuracy, in its gen- outlines, of the theoretical curve of dis- ution of the absolute masses of sediments ss the lateral profile of the oceanic rvoir, as established by N. M. Strakhov. e area of the Javanese island arc, a gen- diminution of the mass of sediment from periphery toward the ocean is observed. ever, as a result of seismic investigations, hickness of the sedimentary layer was d to be non-uniform, and to fluctuate with- range of from 0.4 to 2.5 km.

nalysis of our studies and of published s [5, 8] testifies to the fact that the tec- factor exercises a significant influence the thickness of the ocean-floor deposits. ite the abundant inflow of this material the land, sedimentary material does not mulate in maximum volume on the conti- al slope because of its precipitateness and the great mobility of the water in area. The maximum thickness of the ments was found at the foot of the sub- ne slope and at the bottom of the Bali Trough, which constitutes an enclosed basin.

The steep sides of the mountainous elevations adjacent to the Java Trench are zones of diminished rates of sedimentation, while on the bottom of the Java Trench the thickness of the sediments again increases to 1.5 km.

The influence of the tectonic factor in the vicinity of the Java island arc is also mani- fested in the arrival of sedimentary material in the open ocean. The Bali Trough and the Javanese Trench are sites of abundant con- centration of sedimentary material from the land. In the open ocean sedimentary material comes from the land in small quantities, and there the thickness of the sediments is 0.4 km according to our determinations.

Thus, the nature of the submarine topo- graphy in the region of the Javanese island arc facilitates redistribution of the masses of sedimentary material and the accumulation of sedimentary beds of maximal thickness in tec- tonic troughs constituting contemporary geosyn- clines.

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DOMES CAUSED BY MAGMA BREAKING THROUGH TO THE SURFACE IN LATERAL ERUPTIONS[1]

by

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We know that the ascent of magma to the [su]rface of the earth may be accompanied by [the] formation of uplifts of various amplitudes [and] shapes. There are cases in which the up[lifts] amount to several hundred meters in ele[vat]ion. Their shape may be domed, arched, [or o]f the block type.

Uplifts occur most frequently during the [for]mation of laccoliths and in the bulging of [visc]ous lavas forming the dome. We find ex[amp]les cited in papers by Ye. K. Markhinin, Branca, R. Matteucci, R. Lachman, T. [Min]akami, J. Friedländer, Omori, H. [Ts]otomai, G. Taylor, T. Minakami and K. [...]i, J. Spencer, H. Reck, and J. [...]erzychi [3, 9, 13, 14, 16, 19-23, 25-28,

The uplifts observed by Bemmelen, [...]shkov, Gordon, MacDonald, Eaton, Parsons, [and] Lorenze [1, 6, 17, 18, 25] caused by the [flow]s of lavas of basic and intermediate com[pos]ition are rather rare. Particularly rare are [uplif]ts of the dome type occurring in flows of [...] basalt lavas. The only example of this [kno]wn to us are the domed swells of the [cry]stalline basement beneath certain extinct [vol]canoes in Northeastern Uganda, described [...]erger [5]. In the present article we cite [exa]mples of domed uplifts caused by break[thro]ugh to the surface of basaltic and andesitic-[basa]ltic magma.

From 1954 to 1959 I examined more than [...] adventive cinder cones at the feet of [...]yakskiy, Vilyuchinskiy, Klyuchevskiy, [...]achinskiy, and Ploskiy volcanoes on [Kam]chatka, as well as a number of isolated [cind]er formations in the Topolovaya River [vall]ey, north of Mt. Vilyuchinskiy. Some of [th]e cones are on the arches of domes of oval [or r]ounded horizontal section.

The most clearly pronounced uplift is that of a small cinder cone on the western approaches to Mt. Ploskiy. It is asymmetrical in E-W section and symmetrical in the N-S direction. The angle of slope of the west side is 10-11°, while that of the eastern, northern, and southern sides is 4°-5°, 7°, and 7°, respectively. The uplift disrupts the flat centroclinal dip of the Quaternary rocks on the west and northwest base of Mt. Ploskiy.

A short distance north and south of the cone are the valleys of two streams which form arced bends in opposite directions in the immediate vicinity of the cinder cone (Figure 1).

The formation of the bends was related to a rise in the local uplift in the vicinity of the cinder cone. The gradual rise of this uplift was responsible for additional deepening of the river valleys, change in cross sections and their displacement to the side of the rising uplift.

The formation of the present floodplain and the cessation of lateral erosion testifies to the stabilization of the position of the channel and the absence of further growth of the uplift. The bends in these streams in the vicinity of the cinder cone delineate the northern and southern periclinal termini of the uplift, which, in horizontal section, is an oval with axes of 650 and 500 meters.

The cinders of the cone lie directly on a layer of volcanic ash of Mt. Klyuchevskiy. If we take into consideration the low resistance of this ash to erosive processes and the absence of traces of an overlying stratum of soil, we may assume that the time interval between the formation of the cinder cone and the deposition of its underlying ash layer was quite short. It is possible that the cinder cone was formed against the background of the same eruption of Klyuchevskiy that had laid down the ash.

The Quaternary sedimentation, beyond the limits of the uplift, is horizontal and is overlain, with no traces of interruption whatever, by a pocket of soil-and-ash bed, 0.7 m thick, laid

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[1] [O] kupolovidnykh podnyatiyakh, voznikayush[chikh] v svyazi s proryvom magmy na poverkhnost' [pri b]okovykh izverzheniyakh, (pp. 26-33).

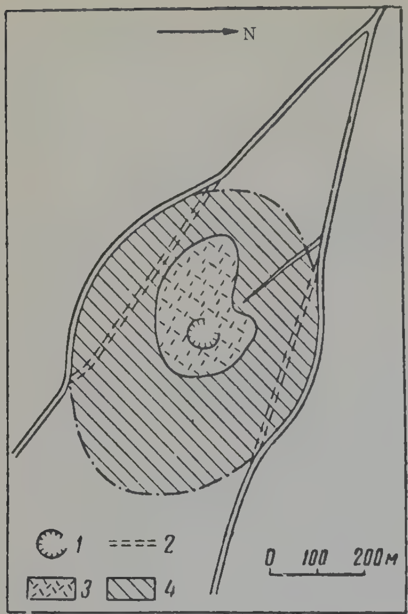

FIGURE 1. Plan View of Dome in Area of Cinder Cone of Mt. Ploskiy

1 - crater; 2 - old stream beds; 3 - cinder cone; 4 - zone of uplift.

down after the formation of the uplift and the cinder cone (Figure 2A), inasmuch as this same pocket fills the drainless crater of the cone. On the periclinals of the uplift, pocket A covers, with an oblique unconformity, pocket B, representing the uplift, while in some places it directly overlies the cinders of the cone.

It follows from the foregoing that the formation of the uplift and of the cinder cone occurred after the deposition of pocket B and prior to that of pocket A. These latter were laid down without interruption and, as a consequence, the domed uplift and cinder cone arose either simultaneously or in direct succession.

Another cinder cone, also on the roof of the dome, lies on the northern slope of Mt. Ploskiy. The existence and nature of the uplift are determined from the corresponding bedding of the soil-and-ash and clayey strata exposed in a number of ravines running radially from the cone.

Comparison of sections of the cinder cone on the periclinals of the uplift and past its limits permits the assertion that the formation of the uplift and the formation of the cinder cone occurred simultaneously. The growth of the uplift was accompanied by the formation of a series of radially oriented fissures, the locations of which are marked by rather deep ravines and runoff channels spreading radially in all directions, including directions opposite to the natural angle of dip. G. S. Gorshkov [1 observed the formation of such fissures in uplifts accompanying the eruption of the adventiv craters on Mt. Klyuchevskiy.

Observations show that in certain instances, clearly defined traces of the existence of radia or radial-and-concentric systems of fissures related to the centers of parasitic eruptions, may serve as reliable proofs of the growth of uplifts occurring in the formation of adventive craters. Two cinder cones may serve as examples of this. One of these is at the western foot of Mt. Ploskiy.

Numerous ravines and small valleys are found in the immediate vicinity of the cone. To gether, they form a radial-and-concentric sys tem, with the cone itself in the center (Figure 3). Stripping performed where the radial ravir approach the cinder cone have shown that, beneath the cinders, the "barrancas" are fissure like in appearance, narrowing at the bottom, 0. 5-0. 6 m in width and not more than 1. 5 m deep. The nearly vertical walls, not having undergone noticeable erosion, are testimony to the fact that the fissures were buried immediat ly after their formation.

To judge by the location of the "barrancas", the center of the dome is beneath the cone. It is obvious that the growth of the dome, causing the radial-and-concentric system of fissures to form, was related to the appearance of a parasitic eruptive center. The existence of the uplift is confirmed by the periclinal bedding of th Quaternary deposits underlying the cone. The diameter of the uplift is approximately 600 meters.

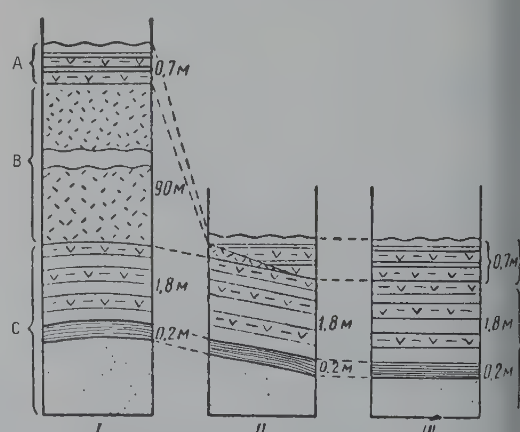

FIGURE 2. Comparison of Sections Through the Cinder Cone on the West Slope of Mt. Ploskiy

I - section through domed portion of uplift; II - section through periclinal uplift; III - section outside of uplift: A - strata laid down after uplift and cinder cone had developed; B - cinder cone; C - strata laid down prior to formation of uplift and cinder cone.

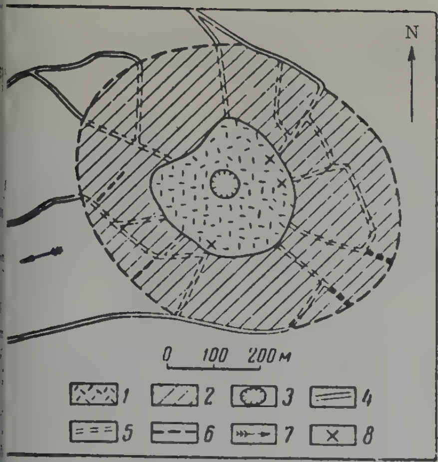

FIGURE 3. Diagram of the Radial-and-
ncentric Fissures in Cinder Cone on the
Western Spur of Mt. Ploskiy

- cinder cone; 2 - dome; 3 - crater; 4 -
all valleys; 5 - same, along fissures; 6 -
pothetical fissures; 7 - slope of surface;
- sites of strippings.

A second cinder cone may be found on the heastern slope of Mt. Ploskiy. The sole in- ation that an uplift occurred here is a rly defined radial-and-concentric network barrancas".

Five kilometers south of this cinder cone e is still another cone. Here the existence n uplift is proved geologically. Moreover, growth of the uplift caused displacement of ver valley not far distant and induced a l change in its longitudinal and lateral files. Displacement of the channel oc- red simultaneously with the growth of the e. The displacement occurred, according ur observations, at the same time as the e rose, inasmuch as the thin pocket of soil- -ash deposits, which in spots covers the ler cone and uplift, is identical to the pocket oil-and-ash deposits covering the gravel in "dry" portion of the valley.

Traces of the uplift on the cone on the south e of Mt. Ploskiy are expressed in the dis- ement of the channel of the stream flowing he western side of the cone. At the site of shift, the valley of the stream is markedly ened, its cross-section is asymmetrical, the stream, forming an arched bend und the uplift, is squeezed against the west- edge of the valley, opposite the cone. With- he limits of the bend where the bed has been placed, there are two clearly marked ter- es, the surfaces of which are folded in an h. The primarily soil-and-ash and sand- -clay layers visible in the sides of the ey form a fold corresponding to the flexure he terraces.

The local nature of the terraces, the fact that they are in the zone of the cinder cone, the arched nature of the terraced surfaces, all are reasons for relating the formation of the terraces to the local dome, which simultaneously shifted the river and produced the bend.

The center of the uplift lay under the cinder cone. The uplift formed in two stages. A flat dome related to the rise of the magma to the surface was formed during the first stage. During the second stage, after some interruption, the activity of the volcano was renewed, and the lava flow broke through to the surface. The uplift caused by the secondary activity of the volcano, was distinguished by smaller scale and only complicated the shape of the dome previously formed and the surface of the upper terrace.

Two cinder cones rise on the southwestern slope of Mt. Ploskiy. The Quaternary deposits underlying the southern cinder cone form an asymmetrical, brachyanticlinal uplift, about 650 to 700 meters long and some 450 meters wide, trending from northeast to southwest. The slope of the southwestern side is 12°-15° while that of the rest does not exceed 5 to 7°. Beyond the bounds of the uplift, the deposits are nearly horizontal, with a barely noticeable slope to the southeast. In the valley of the rivulet crossing the area of the uplift northwest of the cone, at the point of intersection, there is a low terrace (not over 1 meter) the surface of which is folded in a distinctive manner. The highest portion of the terrace is at the point where the valley intersects the axial portion of the uplift. The evolution of the terrace is related to the formation of the uplift.

A horizontal section through the northern cone reveals a rounded flat dome. Domes very similar to those described above exist beneath the cinder cones at the foot of Mt. Viliuchinskiy on the southeastern coast of Kamchatka. Individual elements of the dome-like structure lie along valleys spreading radially from a cinder cone 9 km NNE from Mt. Vilyuchik, on the upper reaches of Lev Topolovyy rivulet. The uplifted area is composed of thin sandy-clayey fluvioglacial deposits covered with a pocket of alluvial sands intercalated with unconsolidated volcanic products.

An uplift embracing the lava flow was observed by us at the cinder cone on the southern slope of Mt. Vilyuchinskiy. The cone is dissected by a deep ravine exposing the upper portion of the underlying foundation, an old lava flow dipping at 6-7°. Directly beneath the cone, the lava forms a gentle fold, the axial portion of which is exactly under the central portion of the cone. The width of the fold along the edge of the ravine does not exceed 60 meters and its length, 200 meters, while in the adjacent parallel "barrancas", the same lava flow is

exposed for a distance of 100 to 150 meters west and east of the cinder cone, but there is no folding.

The lava flow is broken up by numerous fissures, both horizontal (strike azimuth 25°-30°, < 10°-10°) and vertical (strike azimuth 105°-110°, < 70°-75°). Where the lava forms a fold one sees still another system of fissures fanwise to the axis of fold. The origin of these fissures is associated with the flexure of the lava flow and formation of the fold. It would appear that the fold developed as a consequence of lateral volcanic activity, as evidenced by the water-worn cinder cone.

In some instances, the elevated section subsided after uplift. The amplitude of this subsidence may be judged only from the fact that the roof portion of the dome under the cinder cone on the northern slope of Mt. Ploskiy is flattened. This is most probably explained by the pressure of the overlying cinder of the cone. If we consider that, during the initial stage of formation, the roof portion of the uplift had the same radius of curvature as the side, it may be held that the amplitude of subsidence was approximately equal to one-third of the amplitude of the uplift. Judging by the height of the domes, the final amplitude of the uplifts (after subsidence) was between 8 and 30 meters.

All of the uplifts described are flat domes or short brachyanticlines in shape. This gives reason to relate their formation solely to the local effect of vertical forces.

The highest points of the uplift coincide with places at which the magmatic chimney finds outlets at the surface, while the time of formation of the uplifts coincides with the time of formation of the cinder cones, and, consequently, the breakthrough of the magma to the surface, inasmuch as virtually all of these cinder cones are the consequences of volcanic formations resulting from a single event.

It is characteristic that, neither on Kamchatka nor in the Kuriles are domes unrelated to eruptive centers known. The formation of domes beneath volcanic cones due to causes unrelated to volcanic activity, is hardly likely, although R. Bemmelen [7] describes an arching uplift beneath the volcano, Ringit-Beser (Java) which, in his opinion, resulted from horizontal tectonic compression.

The direct relationship, in time and space, between the dome uplifts described and centers of lateral eruptions provides reason to believe that their formation was induced by the pressure of magma and gases rising to the surface. The formation of domed structures caused by the pressure of rising magma was previously indicated by Rech [25] and Bemmelen [8]. It is also clear that a magma breakthrough does not necessarily result in the formation of uplifts. The observations of Fabre [12] may serve as an illustration of this fact.

R. Bemmelen [8] holds that the formation of arched uplifts occurs as a consequence of the filling of a certain amount of space by magma that has risen to the surface, if its volume exceeds that erupted to the surface, regardless of whether the magma rises through a fissure or through undisturbed rocks.

In our opinion, uplifts also may occur when the masses indicated above are equal, inasmuch as the eruptive channels act upon the sedimentary rocks in a manner analogous to the intrusive cores described by Reck [25].

The formation of domes is not likely if the pipe is in a fissure zone, since the chimney then ceases to be cylindrical. During eruptions, uplifts along fissures, if they occur at all, will take the form of arches, the axis following the trend of the fissure. Such upheavals have been adduced by Simotomai [26] and Reck [25].

Aside from the geological environment, the nature of the breakout is also of major significance to the formation of uplifts. If this occurs because of a brief and powerful explosion, and the magma is ejected in the form of semiliquid blocks and flakes, no uplift results. It is specifically in terms of this type of explosion that Fabre [12] explains the lack of traces of mechanical influence of magma upon the rocks which broke through at Mt. Eglazines. If the magma rises slowly or the gas pressure increases gradually, the formation of uplifts will be more likely.

If dome formation has occurred, that in itself is no guarantee that it will be possible subsequently to observe it. The most probably reasons for this are two: subsidence of the uplifted section followed by its burial under flow and ejection materials in subsequent volcanic activity. The amplitude of the uplift attains a maximum immediately prior to the breakthrough of gases and magma to the surface, followed by the onset of the process of subsidence partially, and sometimes, completely compensating for the uplift. Examples of such subsidence are described by Friedländer [14] and Piyp [4].

Subsidences may be associated with posteruptive reduction of the mass remaining in the magma chimney as it cools, as suggested by Büking [11], Knebel [15], or, in the opinion of Lorenzo [17], Branca [10], Williams [29], and others, with subsidence into the emptied pipe. To some degree, the magnitude of subsidence is affected by the pressure of the cinder cone of loose material formed by the eruption. If the volume of ejected products is large enough, the

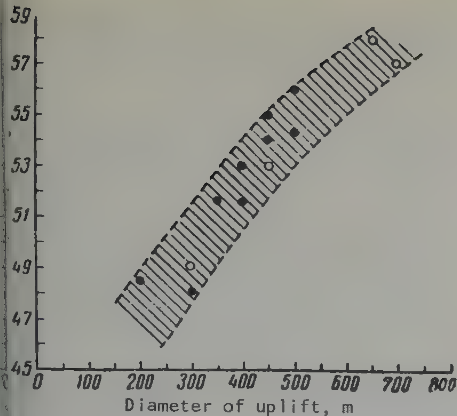

FIGURE 4. Upheaval Diameter Versus SiO_2 Content of Lavas

White circles denote uplifts whose diameters have been determined approximately.

sure upon the elevated section may be so t that the amplitude of subsidence becomes l to the amplitude of the uplift.

here may also be another situation, in h the quantity of loose material is great and rs the entire surface of the uplift, partrly if the latter is small. It is possible that is facilitated by the "creep" of the loose erial on these cones, as has been observed for example, Kulakov [2], and Piyp [4].

he diameter of the uplifts depends upon a ber of factors: the area upon which the ng magma exercises its pressure, the osity of the magma, and the physical proies of the rising rocks. A crude, but unakeable relationship is found between the content (reflecting the degree of viscosity to some degree) in the lavas of cinder cones and the diameter of the corresponding upthrusts (Figure 4).

The conical shape of most of the uplifts we have observed gives reason to assume that the magmatic column whose pressure formed these uplifts had a shape approximating the cylindrical. The shape of a horizontal section through the uplift, usually in the form of a circle or oval, will depend in such a case upon the inclination of the magma pipe near the surface. If, directly at the surface, the pipe is vertical or approximately so, the uplift will be circular in horizontal section, and the resultant fold will have the shape of a regular dome. If the pipe is tilted, the uplift will be oval, and its long axis will be oriented in the direction of the pipe's tilt. The structure is symmetrical in cross section, and asymmetrical along its major axis. The side above the descending pipe will be the flatter (Figure 5).

Of the uplifts we have seen, those which are higher on the slopes of the volcanoes, i. e., where one would naturally assume a steeper location of the feeder pipes, are usually approximately circular in horizontal section, while those at the feet of the volcanoes are more commonly oval, apparently due to a tilting of the pipe.

As the angle of dip of the pipe varies from a right angle to an acute, the shape of the uplift apparently changes from the circular to an increasingly elongated ellipse, and the ratio of the long axis (a) to the short (b) changes from unity (a circle) to infinity. The a-to-b ratio may tentatively serve as a relative index of the angle of dip of the magma pipe at the surface.

The diameter of the uplifts we have described ranges from 300 to 750 meters. Sensitive inclinometers can locate the height of such uplifts

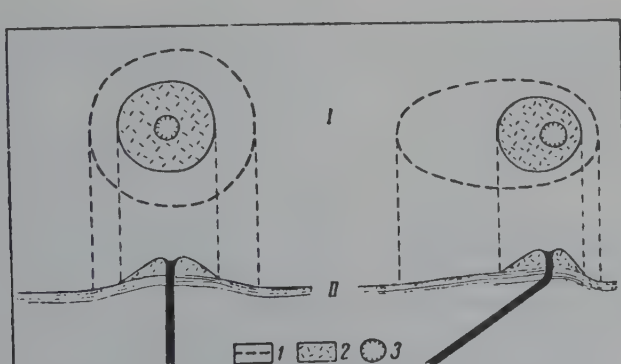

FIGURE 5. Dependence of the Shape of an Uplift in a Horizontal Section and the Nature of Folding on the Inclination of the Feeder Pipe

1 - contour of upheaval; 2 - scoria cone; 3 - crater.

at considerably greater distances from the site of magma emergence. This provides some of the prerequisites for setting up a network of automatic slope measuring stations at the feet of volcanoes revealing a high percentage of parasitic eruptions. The existence of such a network would make possible short-term prediction of at least some portion of the adventive eruptions. On Kamchatka the most likely spot for such a network is the eastern base of Mt. Klyuchevskiy, where there have been seven parasitic eruptions in the past 35 years, resulting in the formation of more than twenty new craters.

In the light of the diameters of the uplifts formed upon magma breakouts, it would appear that a network based on one inclinometer station per 50 to 100 km^2 would be adequate.

Inasmuch as new magma eruptions occur most frequently close to the centers of recent earlier ones, inclinometer stations should preferably be installed at the sites of the most recent adventive craters. Such sites at Mt. Klyuchevskiy would be: the Yubileynaya group of craters, the Bilyukaya craters, the area of Bylinkina crater, and the Tuyla craters.

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THE ELASTIC PROPERTIES OF ROCKS[1]

by

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A knowledge of the elastic properties of rocks is essential to geophysics, geology, mining, and the construction industry. Analysis of these properties also requires knowledge of a number of other properties, primarily porosity and density, and also of the strength constants.

The laboratory of the Institute of Ore Deposit Geology, Petrography, Mineralogy, and Geochemistry of the U. S. S. R. Academy of Sciences has made determinations of elastic constants primarily by the static method (uniaxial compression to one-third or one-half strength), since this method permits analysis of the deformability of rocks by compilation of the appropriate stress-strain diagrams. The dynamic method was used rarely. The Martens instrument was used to measure strain, followed by electrical resistance strain gages. These gages were cemented in pairs at the middle of two opposite sides of a 5 x 5 x 15 cm prism, where uniform stress distributions were assumed. The experimental error in measurements by this method was about 1%. This provided sufficient accuracy in the light of the known non-uniformity of the rocks. According to the data of our experiments, this non-uniformity causes the elasticity and strength figures, even of adjacent samples, to vary by 8-10%.

The stress-strain diagrams showed that a linear relationship is characteristic only of a small number of rocks such as dense quartzites, and gabbro diabases. Other rocks do not follow Hook's Law, and yield curves that are usually concave. The first loading of all rocks having porosity that is at all noticeable shows open hysteresis loops, i. e., residual deformations. Also typical of the first loading is non-uniformity of deformation, which is explainable by non-simultaneous closing of microscopic fissures and pores. On the second loading, the residual deformations disappear in the majority of cases, and the entire measured deformation remains elastic. After long relaxation, the picture may repeat itself, with residual deformations appearing on the first loading. In very weak rocks, residual deformations may be seen on the second and subsequent loadings.

In conjunction with the fact that a knowledge of the general deformation of rocks is required for computations on the behavior of rocks under surface conditions and for engineering calculations, while purely elastic properties are required for analysis of phenomena at depth, we have introduced two indexes: E_1, the "coefficient of deformation", computed from the sum of the elastic and residual deformations, and E — Young's modulus — which is computed solely from elastic deformations. The Poisson coefficients are usually identical for both cases. As a rule, E_1 is smaller than E, but in very dense rocks, where residual deformations are not observed, E_1 may coincide with E even on the first loading.

Theoretically, judging by the values of the coefficients of elasticity and the elastic moduli of crystals (S_{ik} and C_{ik}), the dynamic and static constants should differ from each other but little. The dynamic elastic moduli, adiabatic in nature, have to be only a very little higher than the static, which are isothermal by nature. In the case of metal, this has experimental confirmation. In the case of beryllium copper, according to the data of many laboratories gathered by Richards [12], the difference between E_{st} and E_{dyn} does not exceed 2-2.5%. No such coincidence is found in rocks, as has been shown by Ide [10]; Zisman [16] and others. It would appear that, here, we see consequences of the difference in the physical significance of the effect of porosity in compression and in elastic fluctuations.

The dynamic constants of certain rocks have been determined for comparison on the same prisms as the static. The tests were run simultaneously with O. I. Silayeva (Institute of Terrestrial Physics of the U. S. S. R. Academy of Sciences) on the IKL-5 testing instrument. The constants obtained at identical pressures

[1] Uprugiye svoystva gornykh porod, (pp. 34-41).

ere compared in such fashion that the rates of ropagation of the elastic waves were deterined within the same pressure limits as those ithin which the deformations were measured. the case of dense rocks, the coincidence roved to be satisfactory, but for porous rocks dyn remained somewhat higher than E_{st}. The ynamic Poisson ratios were lower than the atic.

Only completely fresh rocks from the rushers, free of surficial weathering, were sted because, as Table 1 shows, weathering

Table 1

Effect of Weathering on the Physical and Mechanical Properties of Coarse-Grained Microcline Granites of the Novoukraine Type

No.	Rock	Volumetric weight, ρ, $g \cdot cm^{-3}$	Effective Porosity P_e, %	Compressive strength R, $kg \cdot cm^{-2}$	Elastic modulus, $kg \cdot cm^{-2}$	
					E_1	E
1155	Surface-weathered granite	2.54	3.07	1130	$0.90 \cdot 10^5$	$1.61 \cdot 10^5$
1239	Same, weathered to 1.5 m depth	2.61	1.33	1450	$2.34 \cdot 10^5$	$2.94 \cdot 10^5$
840	Same, fissured, from 3-meter depth in quarry	2.69	1.07	1800	$4.39 \cdot 10^5$	$4.70 \cdot 10^5$
830	Same, from 17-meter depth in quarry	2.67	0.8	2390	$4.92 \cdot 10^5$	$5.14 \cdot 10^5$
1156	Same, from 49-meter depth (core)	2.61	0.63	2400	$5.97 \cdot 10^5$	$5.97 \cdot 10^5$

hanges porosity completely and, with it, all her constants of rocks. The elastic modulus particularly sensitive to weathering. From mparison of the E_1 and E values of this table e can also see how the share of residual dermations increases with weathering.

The need to have fresh material limited the lection. Somewhat more than 300 rocks were sted, and their selection proved to be unpresentative: the granites and limestones at are the most widely worked rocks in the S. S. R. were predominant. Complete tables constants have been published in the Trudy stituta geologii rudnykh mestorozhdeniy, trografii, mineralogii i geokhimi of the S. S. R. Academy of Sciences [1]. In the esent article we present a condensed table mean values for 42 groups of rocks selected r geologic and petrographic reasons (Table 2).

The table presents not only the elastic conants but density (ρ) as well as porosity (P) d strength (R).

A dot graph was compiled from the data in able 2, to demonstrate the effect of mineral mposition and porosity upon elasticity. The graph is on a semi-logarithmic scale: E_1 is laid off on the equiscalar ordinate, and porosity on the logarithmic abscissa. The semilogarithmic scale was chosen so as to extend the scale within the low porosity range, where the number of points is large.

The graph illustrates, in highly precise fashion, that the major factor affecting the elastic properties of rocks is its mineral composition and that the effect of factors such as porosity and structure should be considered only within groups of one mineral composition.

Although the effect of mineral composition on elastic properties has already been studied [3, 4, 5, 6, 8, 9, 11], this problem has thus far had inadequate clarification. It is clear from the graph that the mean E_1 for each petrographic group diminishes with increase of porosity in a rectilinear manner, and the position of these straight lines depends on mineral composition. The highest line is that for basic intrusives,[2] while the carbonates are the next lower, then the quartzites and, finally, the granites. Points for individual porous geosynclincal granites, which are indicated by a special symbol, have been introduced for purposes of extrapolation of this last line.

The straight line for the carbonates is in two sections: the points for the geosynclincal carbonates lie on the upper straight line, while the lower, which is steeper, refers to the Paleozoic rocks of the sedimentary mantle of the Russian platform. The last two lines of the table give averaged values separately for the platform and the geosynclinal carbonates and

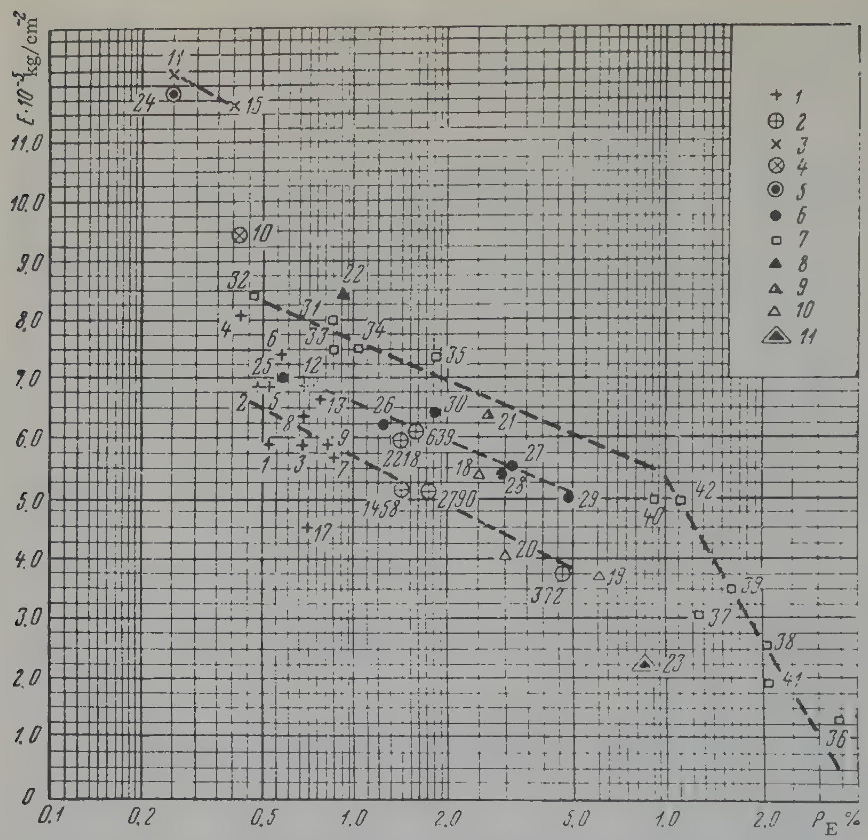
[2] New data show the ultrabasic rocks to lie even higher.

Table 2

Summary of Mean Values of Physical and Mechanical Constants of the Major Groups

No. in diagram	No. of rocks	Groups	Vol. weight ρ, g·cm^{-3}	Porosity P_e, %	Compressive strength, R_c, kg·cm^{-2}	Tensile strength R_t, kg·cm^{-2}	Elastic properties: E_1, 10^{-5} kg·cm^{-2}	E_2, 10^{-5} kg·cm^{-2}	μ	G, 10^{-5} kg·cm^{-2}
		UKRAINIAN SHIELD								
1	3	Migmatites	2.65	0.52	3110	—	5.90	5.95	0.20	2.40
2	8	Plagiogranites, garnet-cordierite granites, charnockites (the oldest)	2.73	0.49	2520	56	6.81	6.86	0.22	2.80
3	15	Gray biotite granites of the 2nd intrusive cycle (medium and coarse-grained)	2.66	0.67	2650	69	5.94	6.04	0,20	2.41
4	2	Fine-grained aplitic granites ("klesovites") of the 3rd cycle of intrusion	2.65	0.42	3600	—	8.10	8.22	0,26	3.34
5	4	Medium-grained Osnitsk biotite granodiorites and granites	2.74	0.52	2550	—	6.80	7.00	0.24	2.85
6	3	Medium-grained Tokovo biotite granites	2.65	0.56	2740	65	7.40	7.22	—	—
7	4	Very coarse-grained biotite granites of trachytoid structure (Novoukraine type)	2.65	0.85	2450	—	5.45	5.49	0.19	2.31
8	3	Fine-grained red granites of the 4th cycle of intrusion	2.64	0.66	2860	—	6.37	6.45	0.19	2.63
9	4	Rapakivi granites of the 5th cycle of intrusion	2.64	0.82	2770	—	5.87	5.92	0.20	2.48
10	3	Coarse-grained labradorites	2.77	0.42	2310	—	—	9.46	0.36	3.48
11	3	Gabbro and diabases	3.00	0.25	3150	—	12.20	12.12	0.33	4.50
		BALTIC SHIELD (KARELIA AND LENINGRAD DISTRICT)								
12	7	Migmatites of the White Sea complex and the Ladoga formation	2.69	0.66	2190	—	6.87	6.92	0.17	2.34
13	17	Medium-grained biotite granites of the 2nd intrusive cycle	2.70	0.78	2440	—	6.72	7.07	0.25	2.46
14	5	Khibiny alkaline intrusives	2.76	—	—	—	5.34	5.62	0.26	2.21
15	2	Gabbro and diabases	3.19	0.40	3200	—	11.70	11.90	—	—
		GEOSYNCLINES								
16	5	Granitoids of the geosynclines (Paleozoic)	2,63	1.96	2510	66	5.61	5.92	0.21	2.63
17	7	Paleozoic granite of the Greater Caucasus Range type	2.66	0.70	2150	49	4.49	4.76	0.22	1.95
18	8	Quartz porphyries	2.59	2.52	2990	—	5.38	5.52	0.17	2.30

19	6	Quartz porphyries of felsitic structure	2.47	6,09	1580	—	3.79	—	0.21	1.56
20	4	Porphyrites of dacitic composition	2.62	3.06	2060	—	4.10	4.38	0.22	1.68
21	3	Porphyrites of Andesitic composition	2.63	2.70	2630	—	6.33	6.48	0.22	2.60
22	2	Augitite porphyrites	2 84	0.94	3645	—	8.34	8.45	0.27	3.28
23	13	Tuffs and tuff breccias	2,49	8.31	900	—	2.27	2.43	0.13	1.00
		QUARTZITES AND SANDSTONES								
24	7	Ferrous quartzites of the Kursk magnetic anomaly	3.51	0,25	3500	—	—	11.9	—	—
25	8	Precambrian and Lower Paleozoic quartzites (Ukraine, Ural, Kareliya)	2.65	0.57	3810	—	7.06	7.14	0.13	3.14
26	6	Quartzitic sandstones	2.65	1.24	2590	—	6.23	6.85	0.10	3.14
27	5	Carboniferous sandstones of the Donbass containing hydromicatized cement	2.63	3.16	2610	—	5.52	5.66	0.14	2.48
28	2	Lower Cretaceous sandstones with carbonate cement (Turkmeniya)	2.60	2.97	2140	—	5.35	5.57	—	—
29	11	Quartzitic Tertiary sandstones of the Volga (opal cement)	2.31	4.75	2260	—	4.94	5.00	0.12	2.23
30	3	Same (chalcedony cement)	2.48	1.79	3050	—	6.41	6.41	0.12	2.91
		CARBONATE ROCKS								
31	13	Fine-grained dolomitic marbles of the Kareliyan ASSR (Kareliya)	2.82	0.83	2790	—	8.07	8.66	0.26	3.12
32	15	Paleozoic marbles (Urals, Central Asia)	2.71	0,46	1520	103	8.38	7.82	—	—
33	4	Marbleized Paleozoic limestones	2.70	0,85	1580	94	7.45	7.22	—	—
34	7	Marbleized Jurassic limestones (Ga. and the Carp.)	2,68	1,05	1550	70	7.52	7.67	—	—
35	3	Mesozoic limestones and dolomites of Turkmeniya.	2,70	1,81	2140	—	7.39	7.77	—	—
36	3	Sarmatian saw limestones of Turkmenia	1.73	34,00	120	—	—	1.10	—	—
37	7	Lower, Paleozoic limestones (Estonia)	2,49	12.37	1060	—	3.17	3 74	0.21	1.56
38	15	Middle Carboniferous of the Moscow syneclise	2.16	21.32	530	—	2.69	2.96	—	—
39	10	Detrital limestones	2.30	15.98	900	—	3.71	3.74	0.25	1.76
40	8	Organogenic limestones and calcarenites	2.57	9.22	1450	73	5.05	5.25	0.32	1.78
41	6	Chemogenic limestones and dolomites	2.14	21.05	420	—	1.86	2.07	—	—
42	19	Mud limestones								
		Limestones & dolomites — Up. Carbonif. —Samara L	2.48	11.12	1350	—	—	5.05	0.23	1.97
		Average for carbonate rocks of the geosynclines (Paleozoic—Mesozoic)	2.70	0.90	1670	—	—	7.70	—	—
		Average for Paleozoic of Russian Platform	2,34	15,65	930	—	—	3.70	—	—



Effect of Mineral Composition and Porosity Upon the Elastic Modulus of Rocks

1 - migmatites and granitoids; 2 - individual granites; 3- gabbro and diabases; 4 - labradorites; 5 - ferrous quartzites; 6 - quartzites and sandstones; 7 - carbonate rocks; 8 - basic effusives; 9 - intermediate effusives; 10 - acid effusives; 11 - tuffs and tuff breccias.

clearly illustrate the significant difference between them: the mean porosity of the platform rocks is 15.6%, while that of the geosynclinal is 0.9%; the elastic modulus for the former is $3.7 \cdot 10^5$ kg/cm2, and for the latter $7.7 \cdot 10^5$ $kg \cdot cm^{-2}$. All this testifies to the effect of metamorphism.

The position of the straight lines on the graph depends upon the elastic constants of the principal rock-forming minerals: quartz, feldspars, pyroxenes and amphiboles, and calcite. The high elastic modulus of ferrous quartzites relative to that of pure quartzites is striking. This is to be explained by the presence of hematite, the mean elasticity of which is $24.0 \cdot 10^5$ $kg \cdot cm^{-2}$ [14].

Inasmuch as the elastic properties of crystals are tensors, the various directions correspond to different coefficients of elasticity (S_{ik}) and moduli of elasticity C_{ik}, derived experimentally by Voigt [15] and, on the average, it is necessary to perform the calculations, with a given margin of error, by means of the appropriate equations of the physics of crystals.

To compute the mean elasticity ($\overline{C}_{ik}$) of crystals of calcite and quartz we used the formulas for averaging proposed by D. Schimozuru [13]:

$$\overline{C}_{11} = \frac{1}{15}(8C_{11} + 3C_{33} + 4C_{13} + 8C_{44});$$

$$\overline{C}_{44} = \frac{1}{30}(7C_{11} - 5C_{12} - 4C_{13} + 2C_{33} + 12C_{44}).$$

If we substitute into these formulas the numerical values of the isothermal elastic moduli C_{ik}, derived experimentally by Voigt [15] and adduced by Shubnikov [7], we obtain the following values of C_{ik} in $kg \cdot cm^{-2} \cdot 10^5$.

For calcite $\overline{C}_{11} = 12.16$ $\overline{C}_{44} = 3.81$

or quartz $\bar{C}_{11} = 10.26$ $\bar{C}_{44} = 4.76$.

y means of Lamé's constants and the known tionships between C_{ik} and E and G [2], we in the isothermic constants.

or calcite $E = 9.69 \cdot 10^5$ kg·cm^{-2} G = $\cdot 10^5$ kg·cm^{-2}.

or quartz $E = 10.16$ kg·cm^{-2} $G = 4.76$ m^{-2}.

he mean values of the Poisson ratios (μ) e computed from the familiar formula $\frac{E}{2G} - 1$. The value of 0.07 obtained for tz indicates its close agreement with the e computed by Voigt — 0.068. The value for calcite (0.27) should be regarded as ewhat understated.

et us now attempt to compare the computed crystals with the parameters of the re- tive aggregates, using pure quartzite and ble of low porosity as the solid quasi-iso- ic bodies, i.e., we arbitrarily disregard possibility that they may be oriented. We that real rocks (the least porous of those have tested) reveal that for marbles E is $\cdot 10^5$ kg·cm^{-2}, while for quartzites it is 10^5 kg·cm^{-2}. In other words, they ap- imate the constants of the minerals, but ys remain somewhat lower because some e volume is pore space.

e are compelled to limit ourselves to this nple of the calculation of E for monomin- ic rocks, inasmuch as the computations exceedingly complex where multi-mineral s are concerned.

he elasticity of quartz, which is low rela- to its hardness, is confirmed by the com- sibility constants (β). A. Adams and E. iamson [8] obtained, for quartz, a com- sibility of $2.60 \cdot 10^{-6}$, while for calcite got $1.39 \cdot 10^{-6}$. The high compressibility uartz is explained by its structure: the tal lattice is of low energy because it has enter of symmetry.

he calculation of E from compressibility ade with the formula: $E = 3K(1-2\mu) =$ $1-2\mu)$, where knowledge of the exact value oisson's ratio is of very great importance If we substitute the values derived for $2.70 \cdot 10^{-6}$, and if the Poisson's ratio is, rding to Voigt, 0.068 (and not 0.25, as h would have it [11]), we obtain, for tz, $E = \frac{3(1 - 2 \cdot 0.68)}{270 \cdot 10^{-6}} = 9.6 \cdot 10^5$ kg·cm^{-2} not 5.12). The value obtained for E coin- s rather closely with that computed above ccordance with the moduli of elasticity: $16 \cdot 10^5$ kg·cm^{-2}. For calcite, with $\mu = 0.27$, we find E to be 9.93, while when $\mu = 0.30$, $E = 8.63 \cdot 10^5$ kg·cm^{-2}. Thus, the value of Poisson's ratio is of great significance in computing Young's modulus from compressibility.

Attention should be given to the Poisson ratios adduced in the tables. In accordance with the exceptionally low values of μ computed for quartz, all essentially quartz rocks have low Poisson's ratios ranging, depending upon composition, from 0.09 to 0.13. Granites and acid effusives also have low Poisson's ratios, at the level of 0.21-0.23. Then follow the carbonates at 0.28-0.30 and the basic rocks at 0.33-0.35.

By way of conclusion we present a table of relationships between constants within the bounds of the major groups, including the propagation rate of longitudinal elastic waves V_{pm} (Table 3).

This table can serve only for the most basic generalizations. To solve specific problems and perform the corresponding calculations, it is necessary to use the detailed tables we published in the Trudy Instituta geologii rudnykh mestorozhdeniy, petrografii, mineralogii i geokhimii Akad. Nauk S. S. S. R., vyp. 43, 1961.

Investigation of the elastic properties of rocks is being carried on in the U. S. S. R. on a fairly large scale in industrial laboratories to solve problems of applied geophysics and mining, for exploratory purposes and for studies of dam foundations. Theoretical studies are also being pursued in the Institute of Terrestrial Physics of the U. S. S. R. Academy of Sciences (Yu. V. Riznichenko, M. P. Volarovich, O. N. Silayeva, O. G. Shamina, and others), in the Mining Institute of the Academy of Sciences, the Mine Survey Institute, and other agencies.

The present study, performed in our laboratory, was possible only because of the friendly cooperation of all of the laboratory personnel, led by Prof. B. V. Zalesskiy. The author regards it as a duty to express his gratitude to all personnel of the laboratory.

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Table 3

Generalized Table of Physical and Mechanical Properties of Rocks

Rocks	No. of blocks	ρ, $g \cdot cm^{-3}$	P, %	R_{comp}, $kg \cdot cm^{-2}$	E, $kg \cdot cm^{-2} \cdot 10^5$	μ	V_{PM}, $km \cdot sec^{-1}$
Basic intrusive rocks	8	2.96	0.35	2850	10.85	0.35	7.400
Granitoids	91	2.67	0.77	2600	6.30	0.21	5.050
Migmatites	6	2.68	0.55	2840	6.75	0.20	5.200
Certain effusives of intermediate and basic composition	8	2.64	2.09	2630	7.20	0.24	5.600
Certain effusives of acid composition	16	3.02	5.21	2695	5.46	0.22	4.500
Ferrous quartzites of Kursk Magnetic Anomaly	7	3.51	0.25	3500	11.90	0.33	7.500
Quartzites and quartzitic sandstones	12	2.65	0.81	3280	7.03	0.13	5.200
Dense sandstones	22	2.44	3.91	2400	5.01	0.13	4.500
Marbles and marbleized limestones, calcitic and dolomitic	37	2.76	0.94	2150	7.85	0.30	6.100
Limestones and dolomites	74	2.35	14.69	1080	3.75	0.24	4.400

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THE PROBLEM OF THE SPECIAL ROLE OF IRON IN THE CRYSTALLIZATION OF SILICATE MELTS UNDER CONDITIONS OF NON-EQUILIBRIUM [1]

by

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Depending upon the conditions of crystallization, multi-component silicate melts of a given chemical composition can yield crystalline products not only of a variety of structures and resulting in mineral species of differing morphology, but having different mineral phase compositions.

In the formation of minerals, these possibilities are manifested most sharply in the crystallization of basic and ultrabasic ferrous silicate melts. The following may be regarded as the principal factors explaining the special role of iron in the non-equilibrium course of the process of mineral formation.

1. The joint appearance of iron in silicates and oxides. When the crystallization process proceeds under non-equilibrium conditions, various shifts are possible in the quantitative relationships between these crystalline phases, depending upon the conditions of mineral formation. As distinct from aluminum, for which this possibility of mineral formation along differing lines is significantly present only in ultrabasic melts, the ferrous melts also reveal significant quantitative changes in the show of iron in the oxide and silicate forms, and in the basic silicate melts.

2. The manifestation of iron in different valences, in accordance with the atmosphere in which the crystallization process occurs, and the speed of that process under non-equilibrium conditions.

3. The dual manifestation of di- and trivalent iron in silicates with coordination numbers of six and four. Despite the fact that, as distinct from aluminum, the coordination number of six is dominant for iron in silicates, it is also possible to find trivalent iron with four as its coordination number in ferric augites, within the Chermak molecule, and in the ferrous variety of gehlenite. Divalent iron in silicates with four as their coordination number exists in the iron variety of åkermanite. The coordination number of iron in the crystal lattices of the silicates is also dependent upon the conditions of mineral formation, particularly when the crystallization process is not in equilibrium.

4. The high development of processes of isomorphism participated in by ferrous silicates and spinelides. This is manifested primarily in the similarity of the structural coordination and the close isomorphic interchangeability of divalent iron and magnesium, as well as of trivalent iron and aluminum. The quantitative and qualitative characteristics of the isomorphic mixtures formed, and the various forms of manifestation of iron in the resultant crystalline phases are determined to a considerable degree by the conditions of mineral formation. The energy relationships of isomorphism, to the role of which attention has recently been directed [6], doubtless testify to the close relationship between the conditions of crystallization and the particular course of mineral formation in the production of simple and complex isomorphous ferrous mineral compounds.

The effects of all of these factors in clarifying the various courses of mineral formation are often very closely related, but when various specific cases of the crystallization of silicate melts are examined, it is often possible to detect the decisive factor in the clarification of the observed variations in phase composition of the rock product.

If we turn to the practices employed in producing silicate rocks under industrial or laboratory conditions (slags, stone castings), in which non-equilibrium mineral-formation processes are most clearly manifest, we will see that the identical melt, of basaltic composition, may yield, depending upon the conditions under which the process of crystallization proceeds, as follows:

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[1] K voprosu ob osoboy roli zheleza pri kristallizatsii silikatnykh rasplavov v neravnovesnykh usloviyakh, (pp. 42-49).

1. Pyroxene plagioclase rocks. The basic plagioclase content of these rocks approximates the theoretical if we assume that the entire calcium aluminate provided in accordance with formula goes to build feldspar.

2. Rocks with a magnetite network structure in which a low-iron calcined magnesia pyroxene is formed. The plagioclase content of these rocks is considerably less than the theoretical.

3. Virtually single-mineral rocks formed by complex ferrous pyroxene. In these rocks plagioclase is either entirely absent or very low in content.

The first type of mineral formation is due to the fact that crystallization is approximately an equilibrium process. This type of mineral formation in basic melts is validated by physical-chemical equilibrium phase diagrams, and, in the case of the basalts, has been analyzed in detail in the petrological constructions of Bowen and Bart. Mineral formation of the second and third types is doubtless associated with non-equilibrium conditions in crystallization, when the degree of deviation from conditions of equilibrium varies.

We have previously [8] directed attention to the shift in the interface between the fields of crystallization of pyroxenes and plagioclases in non-equilibrium processes, and to the special role of aluminum in this phenomenon. In the second and third types of mineral formation we have also encountered a considerable reduction in the field of plagioclase crystallization caused by the formation of complex alumina-bearing pyroxenes and the appearance of a fairly large quantity of aluminum with six as the coordination number. At the same time, in non-equilibrium processes such as these, the special role of iron is manifested no less distinctly in explaining different courses of mineral formation. Specifically, depending upon the conditions of cooling, two non-equilibrium trends are observed in the crystallization of melts containing iron. The first has to do with the crystallization of iron in the oxide form as magnetite, and the second — with the crystallization of iron in silicates in augites-containing iron.

The mineral formation of the former, equilibrium type, is rather well known, and has been reproduced and described in a number of experimental studies. Below we present a chemical (Table 1) characterization of certain synthetic silicate rocks, and indicate the types of mineral formation occurring in each case. It is clear from Table 1 that the silicate rocks described are similar in chemical composition, particularly in terms of acidity.

The conditions of crystallization under which the synthetic silicate rocks were formed were usually determined by the process procedures employed, i. e., rather rapid cooling of a homogeneous melt in the temperature interval of mineral formation (1400-1100°). Some rocks were recrystallized from the supercooled vitreous state at temperatures of 800-1100°.

The following is a petrographic description of the non-equilibrium types of mineral formation.

1. Samples of mineral formation of Type 2 consist of magnetite and pyroxene, with olivine and basic plagioclase present in small amounts. The structural basis of the samples are skeletal magnetite formations (Figures 1 and 2). A special feature of the structure of the majority of the samples is the magnetite network, in the points of which the larger idiomorphic crystals are located. Around the magnetite formations one often observes leached borders and areas demonstrating that the minerals and residual glass around the magnetite are impoverished in iron compounds. All of the remaining minerals except for the olivine are structurally subordinated, in their spatial arrangement, to magnetite formations. The bulk of the basic mass of the rocks consisted of fine crystalline pyroxene, formations of which were grouped in lamellar or spherulite aggregations. Pyroxenes are low in iron, and the refractive indices of 22 samples produced under laboratory conditions (Table 1, analyses 10, 13, 14, 15, 16) fluctuate between 1. 65 and 1. 70.

The refractive indices of certain pyroxenes in rocks produced under industrial conditions (Table 1, analyses 4, 5, and 6) are presented in Table 2.

Olivine crystallizes in skeletal, and frequently spicular crystals and is highly magnesial, judging by the refractive indices (the admixture of isomorphic Fe_2SiO_4 is no more than 7-9%).

Examination of microsections of 24 industrial samples of cast stones having the second type of crystal structure revealed plagioclases only in nine samples, and then only in insignificant amounts (not more than 6 to 10% of the total mass of the stone). The refractive indices testify to the formation of No. 90-96 anorthite. Plagioclase was not found in the copper-nickel slag specimens.

The residual glass seen in alloys of this type of crystalline structure are impoverished in iron and relatively enriched in silica. Thus, the glasses obtained in quenching the melt in an experiment employing charge No. 10 (Table 1) had a refractive index, N, of 1. 650 ± 0. 002, while in an experiment with the 13th charge (Table 1), N was 1. 643 ± 0. 002. In the same experiments, the residual glass in the

Table 1

Chemical Compositions and Types of Mineral Formations of Silicate Rocks[1]

No.	Process procedure	Chemical composition of silicate rocks						Type of mineral formation	
		SiO_2	Al_2O_3	Fe_2O_3	MgO	CaO	R_2O	Second	Third
1	Industrial and laboratory melts of hornblendite	40.91	15.32	16.50	13.30	11.92	1.09	+	+
2		40.04	15.87	15.65	13.88	12.78	1.56	—	+
3		36.51	16.70	21.70	11.95	10.66	1.27	+	—
4	Industrial and laboratory melts of hornblendite with added clay	41.26	14.56	24.64	11.15	9.00	—	+	—
5		40.70	19.60	17.20	10.17	11.52	—	+	+
6		40.60	16.81	20.34	12.04	10.52	—	+	+
7	Slags from nickel and copper melts	42.5	7.2	23.3	8.4	16.0	—	+	+
8		37.84	11.62	28.48	11.11	9.90	—	—	+
9		44.5	11.4	27.8	1.4	9.0	—	—	+
10	Laboratory melts of synthetic mixtures of sedimentary rocks	39.2	13.8	14.1	13.3	20.6	—	+	—
11		42.2	14.9	7.2	13.4	22.3	—	—	+
12		41.8	14.8	17.6	9.3	16.5	—	+	+
13		42.1	14.9	15.9	9.9	17.2	—	+	+
14		42.1	14.8	17.1	9.6	16.6	—	+	—
15		45.9	16.2	15.3	8.0	14.6	—	+	+
16		42.3	14.3	14.5	10.1	15.8	3.0	+	—

[1] The laboratory melts of hornblendite and synthetic mixtures of the sedimentary rocks were made by the author, jointly with L.V. Zverev.

crystalline products had an index of refraction, N, of 1.629-1.633 and N = 1.619-1.622, respectively.

Thus, for silicate rocks with the second type of mineral formation, it is characteristic that a significant portion of the iron separates out as oxide (magnetite), while the silicates (pyroxenes and olivines) are, as a rule, low in iron. A certain differentiation of the oxides in the course of crystallization is observed in the process.

It should be noted that this type of mineral formation, and similar structures, have been described in the experimental studies of I. Morozevich [13], A.S. Ginzberg [4], A.I. Tsvetkov [9], A.A. Leont'yeva [7], and a number of others in which the crystallization of ferrous silicate melts is examined.

2. Samples of mineral formation of the third type are virtually mono-mineralic and consist of iron-bearing pyroxene, which substantially predominates over the other minerals present quantitatively: magnetite, olivine, plagioclase. Ferrous pyroxene is found in thin fibrous, feathery or spherical forms (Figures 3 and 4).

Table 2

Refractive Indices of Pyroxenes and Silicate Rocks With Type Two Mineral Formation (±0.002)

No. of chemical analyses of silicate rocks (Table 1)	1	3	5	6	4
N'_g	1.678	1.675	1.662	1.682	1.697
N'_p	1.653	1.652	1.644	1.658	1.673

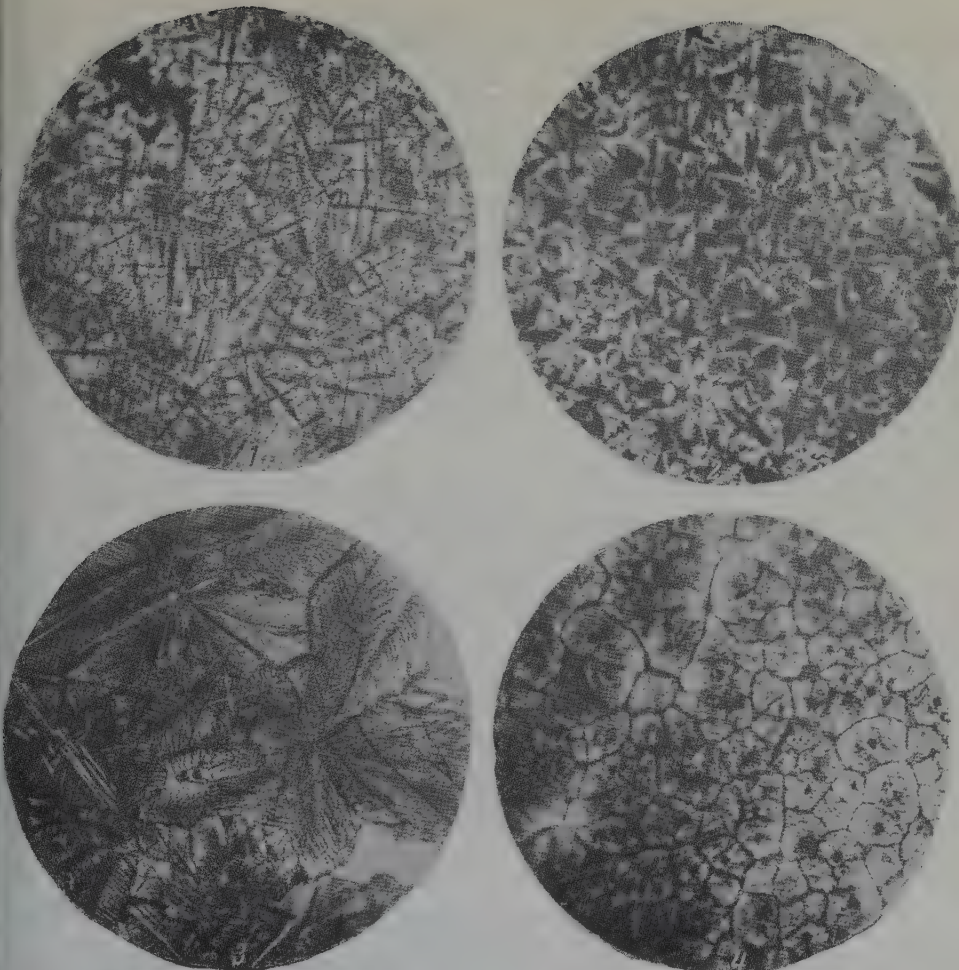

FIGURE 1. Skeletal Network of Crystalline Magnetite Formations.
FIGURE 2. Crystal Structure, Type Two.
FIGURE 3. Crystal Structure, Type Three.
FIGURE 4. Face Structure of Cast Stone of Type Three Mineral Formation.
Without analyzer; magnification 54X.

In some samples we find crystallization of rtually regular polygonal aggregates of lored pyroxene of characteristic rodding ructure with development of spherical pinnate gregates. Identical rodding structure was served by A.S. Ginzberg [3] in cast stones oduced from a melt of Siberian trap. In an amination of the collection of microsections the Moscow Cast Stone Works we encounter- d virtually mono-mineralic cast stone of poly- nal rodding crystalline structure obtained om melts of Onega diabases and Rovno ba- lts.

Table 3 presents the refractive indices of pyroxenes observed in well individualized grains of this type of mineral formation.

A relative increase in the values of the refractive indices of pyroxenes was due chiefly to mono-mineralic rocks, in the formation of which the differentiation of iron oxides characteristic of mineral formation of types one and two was not observed. Complex pyroxenes came into being, incorporating Fe_2O_3 in stoichiometric ratios in Chermak molecules. This conclusion is in accord with the results

of the work of A.I. Tsvetkov [10] in studies of the relationship of optical properties to the composition of pyroxenes in the $CaMgSi_2O_6$ – $CaFe_2SiO_6$ series of solid solutions.

Between the pyroxene aggregates and in the radial intercalations in spherulitic pyroxenes we see residual glass not exceeding 8 to 15% of the total mass in quantity, however. The refractive indices of the residual glass in the thin intercalations between the crystalline formations are very similar to those of glass obtained by quenching the initial melt of crystallization. This last is illustrated in Table 4, compiled from the results of experiments made with a melt whose chemical composition is given under No. 10 in Table 1.

differentiation of rock-forming oxides in type three mineral formation.

A similar type of mineral formation is described in the experimental studies of Dolter [11], Ginzberg [2 and 3], Grigor'yev [5], Tavetkov [10], and Endell [12], and of others, in which the crystallization of ferrous basic silicate melts is examined.

It may be noted that, in a number of silicate rocks, transitional forms of mineral formations between these limiting types are found in spots or in the entire mass, and it is possible to relate the space distribution, the qualitative description, and the relative quantity of the pyroxene formed to the intensity of crystallization and the forms of magnetite growth.

Table 3

Refractive Indices (±0. 002) of Pyroxenes in Silicate Rocks with Type Three Mineral Formation

No. of chemical analyses of silicate rocks (Table 1)	Refractive indexes									
	2	5	5	6	6	8	11	11	11	15
N_g'	1.712	1.682	1.722	1.694	1.718	1 726	1.692	1.720	1.722	1.718
N_p'	1.687	1.660	1.696	1.670	1.692	1.703	1.670	1.695	1.696	1.694

Table 4

Refractive Indices of Glass

No.	Refractive indexes of homogeneous glass due to quenching of melt	Refractive indices of residual glass in intercalations of pyroxene spherulites
1	1,602±0.002	1.603—1.607
2	1,605±0.002	1.602—1.604
3	1.601±0.002	1.603±0.002
4	1.615±0.002	1.617±0.002
5	1.614±0.002	1.612—1.619
6	1.615±0.002	1.613±0.002

The results presented in Table 4 are in accord with D. S. Belyankin's observations of spherulites in technical glasses [1], for which he found similarity in the chemical composition of spherulites and the interspherulite mass. In addition to the mono-mineralic nature of the castings, the similarity noted in the refractive indices of the residual and initial glasses for crystallization confirms the lack of significant

The examples presented reveal the varieties of behavior of iron when mineral formation occurs under non-equilibrium conditions. The large percentage of magnetite with most of the iron present as oxide, the formation of virtually mono-mineralic rocks comprised of ferrous augite, the crystallization of magnesial olivines from melts containing considerable quantities of divalent iron, the substantial shift in the boundary plane between the fields of crystallization of pyroxenes and plagioclases, the varying degrees of differentiation of the rock-forming oxides – all of these are indices of varying degrees of a non-equilibrium course in the process of crystallization of a ferrous melt.

Thus, in examining the question of the mechanism of type two mineral formation and its structural basis – the magnetite skeleton structure – it may be seen that the ratio of ferric to ferrous iron increases noticeably in crystalline specimens as compared to the same in the homogeneous initial melt for crystallization. According to the data of chemical analyses of samples of glasses obtained in quenching homogeneous melts, the Fe_2O_3/FeO ratios, established as the stone becomes cast, are quite constant, and amount to 0. 70-0. 74. At the same time, in finished castings of

mineral formations of type two, these ratios are always of higher values and fluctuate in the range from 0.83 to 1.18 (from 19 analyses of various castings).

The comparatively rapid cooling of the melt tended to result in intense magnetite formation in many centers of crystallization because of the exothermic process of oxidation of divalent iron. Equilibrium crystallization mechanisms would indicate that the magnetite separating out should react with the surrounding silicate melt, with consequent formation of ferruginous silicates. However, the high rate of cooling tends to produce rapid loss of heat from the spheres of these reactions while the rapid increase, under these conditions, of the viscosity of the residual melt, governs, as it were, the formation of magnetitic skeletal networks which, in the form of thin, complex dendrites, penetrate the entire mass of crystallized silicate melt. Meanwhile, the relatively smaller and more mobile ions of iron diffuse, within given limits, into the growing crystals. This is indicated by the leached sections and the borders around the magnetite crystals. Microscopic investigation of a number of silicate rocks having this type of crystalline structure revealed crystallization of magnesial olivine in quantities exceeding its computed theoretical molecular ratios. This is also explained by violation of the equilibrium course of the process of crystallization in rapid cooling of the melt. After the iron fixed in the magnetite is removed from the sphere of possible reactions, relative enrichment of the residual melt in calcium, magnesium, aluminum, and silicon occurs. At the same time, the specific effect of the iron ceases to make itself felt. This is expressed in reduction of the temperature of crystallization of the possible ferruginous silicates. Thus, because of both the high rate of cooling and of specific changes in the chemical composition, the melt remaining after formation of the magnetite networks and the precipitation of the olivine, remained in a condition of considerable supercooling. Under these conditions one observes rapid crystallization of the pyroxenes, which are of the calcium magnesia type, on the basis of their optical properties.

A significant factor in mineral formation of type two is a crystallization sequence in which magnetite separates either before, or simultaneously with, the magnesial olivine. This latter is possible if the melt contains not less than 12%-15% iron oxides. Under these conditions, the temperature at which the crystallization of the anticipated ferruginous silicates in a multi-component melt starts is lower than that for crystallization of magnetite.[2] All other conditions being equal, the formation of this type of crystal structure is facilitated if the initial melt is of low viscosity (low acidity and higher content of iron oxides). In the process of development of the magnetite network, viscosity increases markedly both in the heterogeneous system as a whole and in that of the residual melt.

We related the development of the nearby mono-mineral silicate stones with an even larger deviation from a state of equilibrium during the crystallization process. Under these conditions, differentiation of the rock-forming oxides occurs to a small degree in mineral formation, and the field of crystallization diminishes to the extreme because of the aluminiferous and ferruginous augites. In other words, silicates of more complex crystalline structure are not formed from the relatively simpler silicates, however, trivalent iron is manifested to a considerable degree with a coördination number of four, which is not typical of it.[3]

This type of mineral formation can be reproduced particularly well with this type of cooling of the melt if crystallization of the casting occurs "from below , i.e., when pronounced supercooling results in the formation of a primarily vitreous mass which then recrystallizes because it is held at a given temperature, or supplementary heating is applied. The appearance of mono-mineral silicate rocks of this type is most frequently observed in the crystallization of the more acid melts (SiO_2 content up to 50-52%), in which pronounced supercooling is facilitated by relatively rapidly increasing viscosity.

We thus come to the conclusion that, in an examination of questions of the genesis of basalts and the conditions of production of synthetic rocks of basic and ultrabasic compositions, it is possible to arrive at specific results by examination of the various forms in which iron manifests itself in the mineral phases.

The particular role of iron is its exceptional sensitivity to various deviations from an equilibrium course in the crystallization process, when various ferruginous mineral phases occur only within comparatively narrow and special conditions of mineral formation.

The equilibrium mechanisms of crystallization of basaltic magma have been elaborated in

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[2] The temperature at which magnetite crystallization starts was determined by us for a Ural hornblendite melt to be 1250°-1265°. A. A. Lenont'yeva [7] found it to be 1300° for Transbaykal basalt, and A.I. Tsvetkov [9] found that for an Onega diabase melt it was 1260°.

[3] The ratio of the ionic radius of trivalent iron to that of oxygen is 0.51. This ratio lies within the limits of stability of coordination polyhedra with six oxygen atoms surrounding a metal atom (ratio in 0.41-0.73 interval). The possibility that Fe^{+3} can replace Si^{+4} with four as coordination number is much smaller than that Al may do so, because the ratio of the ionic radius of aluminum to that of oxygen is 0.43 — a value approaching the limit of stability of structures having coordination numbers of four and six.

petrological theory to a high degree of reliability, and therefore the deviations from these mechanisms or from the corresponding theoretical molecular ratios observed in study of the final crystalline products may, particularly in the study of ferruginous minerals, reveal the specific features of formation of a given mineral rock.

The most significant factors are the degree of differentiation of rock-forming oxides, the shift of the boundary plane between the fields of crystallization of pyroxenes and platioclases, the quantitative distribution of the iron between the oxides and the silicates, the nature of the manifestation of iron in isomorphic mineral compounds — olivines and pyroxenes, the manifestation of iron with four as coordination number in the silicates, and the iron content of the residual glass.

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EPIGENE ZONING OF URANIUM MINERALIZATION IN PETROLIFEROUS CARBONATE ROCKS[1]

by

V. N. Kholodov, et al.

INTRODUCTION

As we know, there is a group of uranium deposits in which mineralization is paragenetically related to organic matter of the petroleum series. This type of uranium deposit still has not been adequately studied, but it is doubtless of major interest both from the theoretical and practical points of view.

In the same category as deposits associated with petroliferous bitumens are ore manifestations in carbonate rocks, and the purpose of the present article is to provide a generalized description of these.

It has long since been established that this type of mineralization usually tends to appear in specific horizons. This stratigraphic pattern in the mineralization is deemed to be an argument in favor of the syngenetic origin of uranium mineralization. On the one hand, we know that carbonate strata undergo significant changes in facies. However, the ore therein is localized in narrow bands, the position of which is not governed by these changes. This circumstance would tend to indicate that the mineralization is epigenetic.

The data accumulated in the study of various deposits and discussed in this article make it possible to speak with greater certainty of the origin of this type of mineralization and to regard it as the result of epigenetic processes.

ZONAL STRUCTURE OF ORE-BEARING HORIZONS

The ore-bearing horizons of uranium deposits associated with petroliferous carbonate rocks frequently display secondary zoning. This phenomenon is so widespread that it can be used as a distinctive sign of ore in prospecting. Figure 1 presents a generalized diagram of the epigenetic zoning of an ore-bearing carbonate horizon. The upper portion of the graph is a lithological profile of the facies. Five lithologically distinct strata may be seen in its most complete section.

Layer 1. At the base there is a limestone dolomite containing numerous fossils of small pelecypods, spottily impregnated with brown bitumen. The layer, can be easily followed over a considerable area and can serve as a marker horizon.

Layer 2. Overlying the former is a limestone of organogenic structure consisting chiefly of the shells of small gastropods and foraminifera, with a pronounced dominance of miliolites. The fossil fauna are usually cemented by a calcite of medium crystalline structure. As one proceeds to the top of the stratum, the rock begins to show limestone oolites. The transition to the overlying layer is therefore gradual.

Layer 3 is represented by limestone of large oolitic structure, very homogeneous in thickness and along the strike. The limestone oolites are cemented by coarse crystalline calcite. The virtually complete absence of fauna remains is noteworthy.

Layer 4. Dolomitic limestone, structurally similar to the preceding layer, differs from it significantly in that it contains calcitized small pelecypods, comprising, with the oolites, the bulk of the rock. Dolomitization, which increases regularly toward the top, is characteristic. Dolomite rhombohedra replace the oolite nuclei, edge the fossil fauna and are distributed among the calcite grains of the cement.

Layer 5. The section is completed with a stratum of pelitomorphic limestone dolomite containing inclusions of fossil fauna and oolites filled with coarse crystalline calcite.

In chemical composition and the manifestations of secondary calcitization, the rocks of

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[1] Ob epigeneticheskoy zonal'nosti uranovogo [...]deneniya v neftenosnykh karbonatnykh porodakh, [...], 50-63).

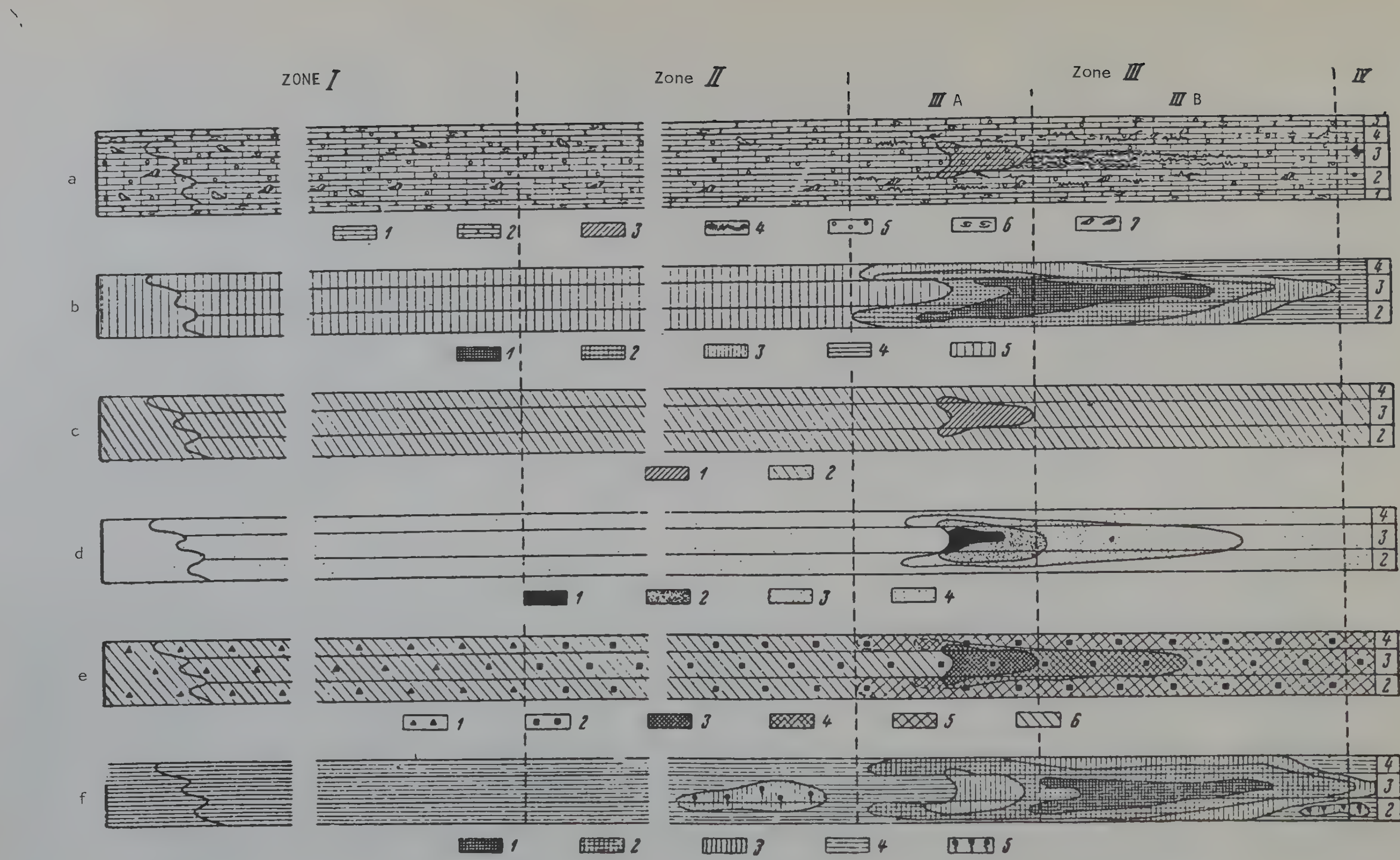

FIGURE 1. Epigene Zoning of the Ore-Bearing Horizon.

a - lithologic-facies profile: 1 - limestones; 2 - limestone dolomite rocks; 3 - silicaon concretion; 4 - sutured stylolite joints; 5 - oolites; 6 - pelecypod shells; 7 - gastropods; b - uranium distribution: 1-5 - uranium concentration in rocks, in order of declining magnitude; c - silica distribution (%): 1 - 25-50, 2 - traces; d - % distribution of vanadium; 1-0, n, 2 - 00n, 3 - 0 00 n, 4 - 0 000 n. e - distribution of (total) iron: 1 - region of Fe hydroxides; 2 - region of Fe sulfides; 3-6 - total Fe content, %; 3 > 0.60; 4 - 0.60 - 0.40; 5 - 0.40 - 0.25; 6 - 0.25-0.10; f - distribution of organic carbon (total, %): 1 - 0.60 - 0.40; 2 - 0.40 - 0.20; 3 - 0.20-0.10; 4 < 0.10 5 - impregnated with liquid

Strata 1 and 5 are very similar. Throughout the profile studied they contain no commercial mineralization. However, the limestones of layers 2, 3, and 4 contain extensive uranium mineralization within the limits of the deposit.

The type of section described is essentially the same over virtually the entire length of the profile, except that in the vicinity of the shoreline of the ancient basin (the left side of the figure) the facies of layers 2, 3, and 4 are replaced by a non-stratified layer of limestones of organogenic-oolitic structure, with an admixture of fragmented material.

Cherty limestones are prominent in one section of the profile. They form concretionary bodies essentially conformable with the stratification of enclosing rocks in the form of an elongated half-moon in the section. Immediately adjacent to the siliceous section is a whole system of large stylolitic joints, also essentially concordant with the stratification.

The ore-bearing horizon was sampled and described in detail over its entire length, with chemical determination of the SiO_2, Fe, and C_{org} content in the samples. S semiquantitative spectral analysis was used to determine the vanadium content, while the uranium was determined by radiometric and chemical analyses. The results are presented in Figures 1b, c, d, e, and f.

Comparison of all the diagrams of Figure 1 permits us to identify, in the ore-bearing horizon, four lithologic-geochemical zones, characterized by different relationships of the elements determined and by secondary specific changes in the enclosing rocks.

Zone I is distinguished by extreme facies non-homogeneity, since it includes the non-stratified terrigenous carbonate rocks of the ancient coast, as well as the bedded organogenic oolitic limestones and dolomites (Figure 1); takes its origin from outcrops of the carbonate level at the surface and may be followed at depth to great distances along the dip of the rock.

The basis of the geochemical distinction of the zone is the wide dissemination of iron hydroxides. Everywhere, they fill thin fissures, cavities and porous sections of the rock, giving it a spotted yellow-brown tint. Characteristic is the virtually complete absence of uranium in the rocks ($8 \cdot 10^{-5} - 2 \cdot 10^{-4}\%$), the small quantities of vanadium (reaching $n \cdot 10^{-4}\%$), and the negligible amount of organic carbon (not over 0.10%). Highly developed porosity and vesicularity is characteristic of all the various carbonate rocks. Most frequently the pores and cavities are developed as a consequence of the leaching of the cement, the limestone oolites or shells. Hollows are formed in the oolites and shells, and the rock as a whole takes on a spongy structure.

As may be seen from Figure 1e, the iron hydroxides, are replaced at some distance by pyrite and marcasite. The bottom of zone 1 is drawn at the point of demarcation of the ferric and ferrous forms.

Zone II is bounded on one side by the oxidation-reduction boundary for iron, shown in conventional symbols in Figure 1e, and on the other by the boundary between the ore-containing and the ore-free rocks. The iron content fluctuates between 0.1 and 0.25%, and most of it is in the form of pyrite and marcasite. The fresh and clearly defined crystals of pyrite stand out in the pores of the carbonate rocks. In spots the coarse crystalline pyrites are fringed by selvages of calcite veinlets intersecting the bedding (Figure 2). These features of the bedding of the iron sulfides are testimony to their late, epigenetic, origin.

A second and very important characteristic of Zone II is the virtually complete absence of organic matter, expressed in the marked leaching of the carbonate rocks. One finds only isolated spots of brown-buff bitumens concentrated in the most porous sections. These are clearly of a later origin than the leaching processes. Within the bounds of the zone studied, the uranium and vanadium contents do not exceed $3 \cdot 10^{-4}\%$ uranium and $n \cdot 10^{-4}\%$ vanadium.

The carbonate rocks of this zone, similar in facies to the rocks of the adjacent lithological and geochemical zones, differ from these in that they show traces of intensive solution and secondary recrystallization, both equally embracing the entire thickness of the section.

When the rocks are studied beneath the microscope, one may see here the solution and displacement of individual parts of the oolites and shells and, as a consequence, the formation of brecciated structures and thin branching, solution fissures (Figure 3a). It is of interest to note that simultaneously with the process of solution, one observes in the rocks of this zone an intensive secondary calcitization, the coarse crystalline calcite replacing oolites, shells, and other formations (Figure 3b).

Thus, Zone II as a whole is highly reworked by secondary processes. This is evidenced by the epigenetic nature of the pyrite, the universal removal of organic matter from various layers of the section, intensive solution, and subsequent calcitization of the carbonate rocks. Its width is only a fraction of that of Zone I.

The general outline of Zone III coincides with the uranium mineralization.

A section through the ore body frequently

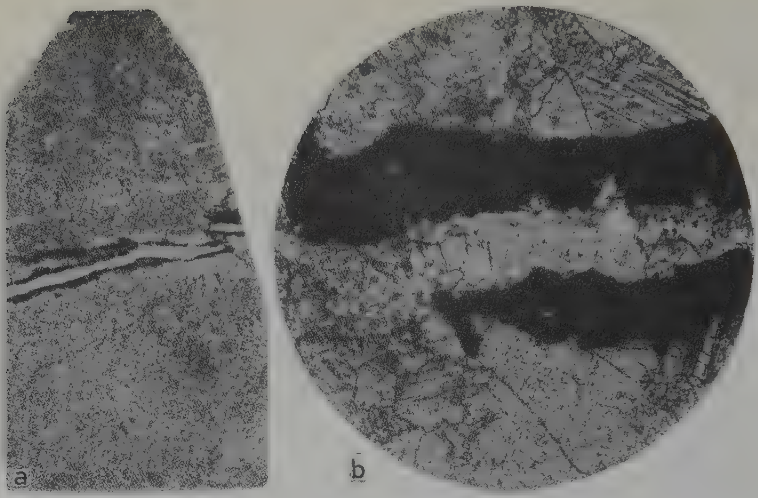

FIGURE 2. Epigenetic Formations of Pyrite in Selvages of Calcite Veins

a - sample of oolitic limestone. Actual size; b - same sample (photomicrograph). 63X.

shows it be a half-moon in shape, markedly elongated along the bedding of the rocks (Figure 1b). In a number of sections, its boundaries intersect the layering of the enclosing limestones, as a consequence of which uranium mineralization is encountered in layers 2, 3, and 4. It is characteristic that a marked transition from the ore-containing to the non-ore-containing rocks is observed in the concave portion of the ore body coinciding with the silicon concretion. Its boundary is highly blurred on the opposite side, and here one sees a very gradual and regular diminution in uranium content.

The pronounced asymmetry, the sickle shape, and the intersection of the contours of the ore body by bedding planes of the enclosing rocks is testimony to its secondary, epigenetic, origin.

Ore bodies highly similar in morphology are known in the geological literature as "rolls". The majority of investigators are also inclined to regard them as epigenetic in origin [7, 20, and others].

From the point of view of the primary facies peculiarities of the enclosing rocks, the ore zone differs in no way from non-ore Zones II and IV. None of the three beds participating in the structure of the ore body undergoes significant changes over the entire extent of the profile. As far as the secondary features of the intrusives are concerned — these are highly significant and permit the identification of two sub-zones. On the left side of the ore body there are siliceous ore-bearing rocks (Subzone IIIA). The right side is represented by ore-bearing limestones cut by numerous stylolite joints (Subzone IIIB).

Subzone IIIA. The silicon concretion in the left side of the ore body repeats its configuration in miniature. Its major portion lies within layer 3, and two subordinate processes enter layers 2 and 4. Thus, like the ore body, the silicon lens is shaped like an elongated crescent moon in section.

It is characteristic that the layers of carbonate rock may be followed without significant changes into the silicon concretions, and that their boundaries are not bent. As we know, such relations indicate that the concretion came into being in rock already formed.

Along the concave portion of the silicon concretion, the contact is extra-ordinarily clear-cut and even. Here we observe an accumulation of viscous black bitumen that marks the contours of the silicon concretion in the form of a film, 5 to 6 mm thick. Within the siliceous body, the quantity of organic carbon is considerably reduced relative to the enclosing rocks (Figure 1f). This distribution of organic matter is induced, in all probability, by the silicification of the carbonate rocks, in the process of which the bitumen initially contained therein was forced out to the periphery of the concretion.

In the concave portion of the siliceous body one also finds another type of contact. The

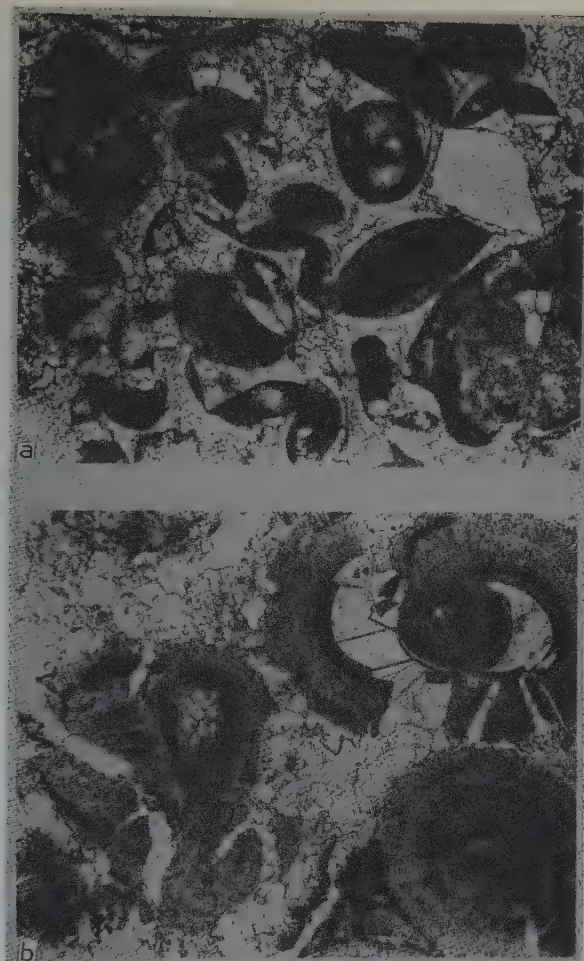

FIGURE 3. Manifestations of the Processes of Solution in Calcitization in Rocks of Zone II

a - solution and displacement of oolites in oolite limestone, with formation of breccia-type structure; b - secondary calcitization in oolite limestone. Magnitication 37X.
Magnification 37X.

interface between the silicified and non-silicified limestones is diffuse, and organic matter is present here in small quantities.

As indicated in Figures 1b and 1f, the maximum concentration of uranium, like that of organic carbon, tends to be in the convex portion of the concretion and is pushed beyond its limits. Conversely, the maximum vanadium and iron contents are within the concretion (Figures 1d and 1e).

The structural and textural features of the silicon carbonate rocks forming the interior portions of the concretions differ little from the limestones enclosing them. Quartz and chalcedony replace certain structural components of the rock in such a way that they retain their initial appearance. Various cases of substitution are illustrated in Figure 4. One sees clearly that the oolites, the pelecypod shells, the gastropods and the sections of cement are silicified without disturbance to their sculptured shapes. It is quite characteristic that large quartz crystals were formed where one might assume the original presence of coarse crystalline calcite. A similar regularity has been described in a paper by S. G. Vishnyakov [2], who holds that "the silica has replaced a precipitate in which all vesicles had been filled by crystallized calcite, i.e., a precipitate that had converted to rock."

Curious results were obtained in a study of the effective porosity of siliceous limestones in layer 3 (Table 1) using the Preobrazhenskiy method [11].

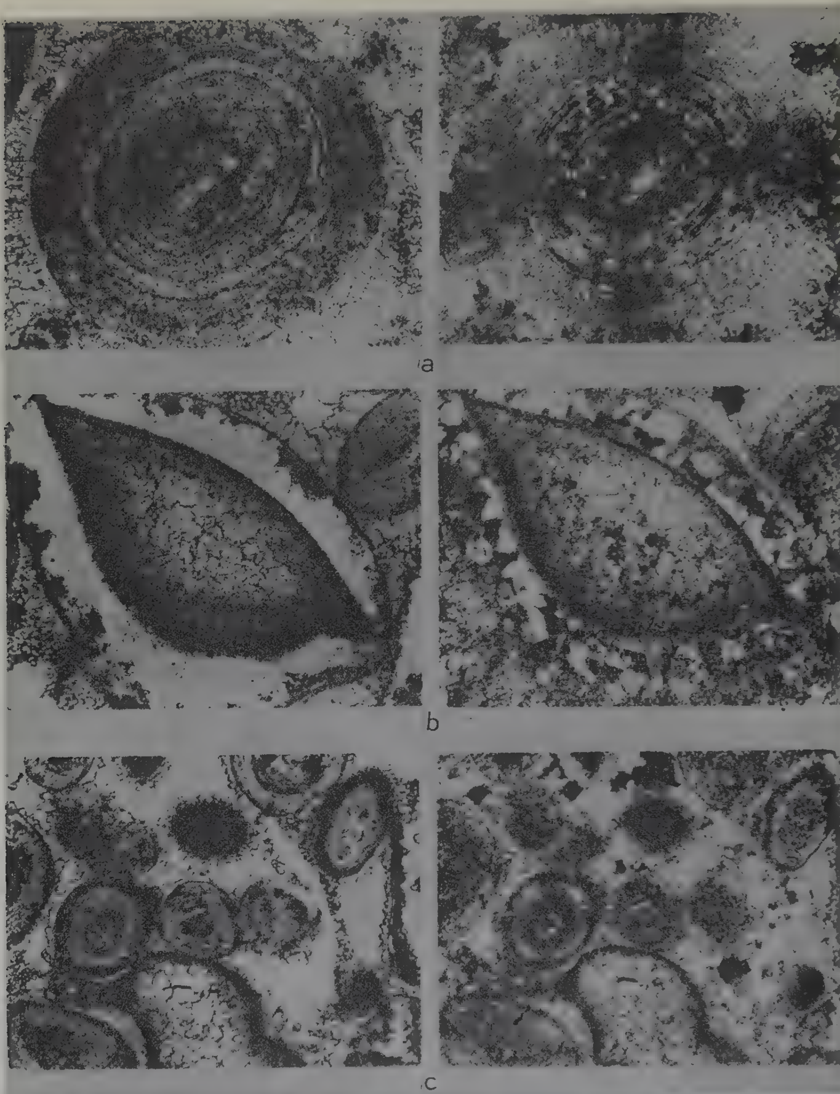

FIGURE 4. Replacement of Various Structural Elements of Limestone by Silicon

Left: photomicrograph with a single Nicol prism; right; with crossed Nicols. a - silicification of oolite, magnification 70X; b - replacement by quartz of pelecypod shells, magnification 70X; c - silicification of cement of oolite limestone, magnification 35X.

It follows from the data in Table 1 that the effective porosity of siliceous rocks is only 1/5.5 as great as in nonsiliceous rocks. Consequently, the process of silicification occurred in this case, not only by substitution of various parts of the carbonate rocks, but by filling of the fissures, pores, and vesicles.

Consideration of all the foregoing characteristics of the siliceous concretions studied permits the conclusion that they were formed in the epigenetic stage.

Subzone IIIB is that portion of the ore body that was not affected by the processes of secondary silicification. The maximum concentrations of uranium are close to the boundaries of the silicon concretion and coincide in space with the zone of extensive development of the stylolite joints.

Uranium minerals in association with organic matter and iron sulfides fill the cracks and pores of the carbonate rocks, are concentrated in the sutured stylolite joints, are encountered

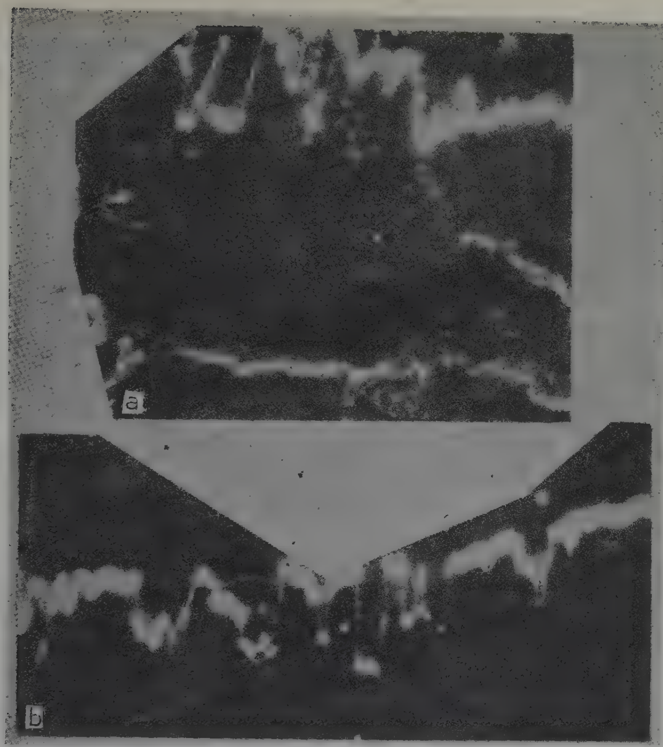

FIGURE 5. Uranium Mineralization in Sutured Stylolite Joints

a - dense network of stylolite joints at edges of silicon lens; b - stylolite joint parallel to bedding of rocks in center of ore body.

in the cement and the oolites, and fill in the casts of the gastropod and pelecypod shells, and other fossil fauna.

Macroscopically visible uranium minerals are extremely rare. Special mineralogical researches have made it possible to find submicroscopic secretions of pitchblende and black oxide of uranium. Moreover, the existence of organic uranium compounds is hypothesized (I. G. Chentsov).

Uranium in the fissures and pores of ore-bearing limestones is found by comparing radiographs and photographs of polished samples, as well as by examination of sections under the microscope. Finds of this type are widely disseminated throughout the entire ore body. They testify to the fact that the process of ore formation occurred in rock already formed, in which these textured features were developed.

Uranium in stylolitic joints, highly developed over the entire area of the ore body, is the major form in which this element is found. This is clearly evident from comparison of curves a and b in Figure 1. Toward the periphery of the ore body, the quantity and dimensions of the stylolites gradually diminish, and the boundaries of the ore body are essentially determined by the limits of distribution of these formations.

In the center of the ore body, corresponding to layer No. 3, one usually finds a number of large stylolite joints, approximately parallel to the bedding. The columns of these attain 2 to 3 cm in height. At the boundaries of the lenticular belt, the stylolite joints branch and intersect the inclined stylolites to form a dense network. As the distance from the silicon concretion increases, the dimensions of the columns of the stylolite joints diminish. They combine, taper off, and become fine fissures. Small sutures and microstylolites are highly developed in the border zone of the ore body. In the majority of cases these intersect the bedding.

All of the suture stylolite formations contain a clayey film constituting an insoluble residue of the enclosing rock. Here, too, we

Table 1

Effective Porosity of Silicified Rocks (For Purposes of Comparison the Same Data are Presented for Non-Silica Rocks of the Same Stratum)

Effective porosity of silicified carbonate rock, %		Effective porosity of non-silica carbonate rocks, %	
In samples	Mean value	In samples	Mean value
3.37	2.80	19.50	15.24
2.90		15.88	
2.01		12.15	
1.76		14.60	
5.48		15.95	
2.14		16.50	
1.96		14.80	
		17.95	
		16.31	
		10.37	
		13.52	

find organic matter, pyrite, and uranium minerals (Figure 5).

There is an extensive literature on the genesis of sutured stylolite textures. However, the origin of these formations has been envisaged from different points of view. The most favored hypothesis holds that stylolite joints are the result of solution of carbonate rocks at point of maximum pressure. Convincing proofs of the secondary origin of stylolite joints are presented in the work of Blake and Roy [22], Rigby [25], Uspenskiy [16], Dunnington [23], and Kholodov [18]. In his most recent work, Teodorovich [14] also recognizes the possibility that stylolite joints and sutures are of epigenetic origin.

The sutured stylolite joints of Subzone IIIB, like many stylolites, are characterized by numerous signs of secondary origin. Among these one may note widely distributed tilted and intersecting stylolites in a close relationship between these and fissures, association with breccia zones, etc. This compels us to assume an epigenetic origin of the uranium concentrations associated with the stylolite joints.

Uranium in oolites. Punctate concentrations of uranium in the internal portions of limestone oolites are widely developed in ores.

It is found under the microscope that mineralization is found in all oolites regardless of their state of preservation. Organic matter and iron sulfides in association with uranium minerals are arranged along narrow radial fissures and in the interior cavities of oolites, and are also distributed through the individual concentric shells, or permeate them in their entirety.

The means by which uranium concentrations occur in limestone oolites is determined by the permeability of their outer layers.

Until recently oolites were deemed to be virtually impermeable and, therefore, the petroliferous bitumens they contained were regarded as syngenetic, i. e., as having originated simultaneously with the oolites.

One of our studies [19] presented data testifying to the high permeability of the outer layers of limestone oolites. Thus, uranium concentrations within oolites may theoretically appear at any stage of rock-formation, including the epigenetic.

Uranium in fossil remains. In the rocks of layers 2 and 4 we find, along with mineralized oolites, the shells of gastropoda and pelecypoda, containing uranium minerals. The latter either fill the interior cavities of the shells or partially replace the calcite forming their walls. Phenomena of secondary replacement of the shells by quartz, chalcedony, and dolomite, as well as cases of the filing of the casts of fossil fauna with secondary bitumen, which are frequently observed in the same sections of the stratum, testify to the possibility of formation of similar concentrations of uranium in the epigenetic stage.

If we examine the various forms in which uranium is found within the bounds of Subzone IIIB, it is necessary to emphasize, in general, that the maximum concentrations of this element, and apparently the greatest portion, is related to stylolite joints, pores, and fissures.

In the ore bodies studied, isolated uranium minerals are rarely encountered. The bulk of the element is usually closely related to the organic matter, the iron sulfides, and to clayey minerals, and forms an ore complex with them. Along with uranium, the ore complex reveals increased amounts of vanadium, cobalt, nickel, and molybdenum. The ore complex fills the pores and fissures, impregnates the oolites and shell casts and concentrates with particular frequency along the planes of the sutured stylolite joints, thus filling the most premeable portions of the rocks. It would be natural to anticipate that the effective porosity of the ore-bearing portions of the bed will be less than that of the non-ore-bearing.

The results of a comparative study of porosity are presented in Table 2.

It follows from Table 2 that the porosity of the rocks in ore-bearing sections is diminished

Table 2

Effective Porosity of Barren and Ore-Bearing Carbonate Rocks

Stratum	Barren rocks of Zone II		Metalliferous rocks-Zone III		Barren rocks of Zone IV	
	Eff. por. % (samples)	Average for stratum	Eff. por. % (samples)	Average for stratum	Eff. por. % (samples)	Average for stratum
4	— — — 12.05 5.20	8.70	1.62 1.53 4.02 4.00 6.07	3.45	11.15 5.90 10.90 — —	9,32
3	14.80 16.50 12.15 19.50 15.88 — — —	15.77	14.60 10.37 15.95 5.75 17.95 13.52 16.31 17.25	13.96	18.20 16.60 15.00 17.58 14.20 — — —	16.32
2	— 7.30 12.30 — — —	9.80	3.94 6.90 3.15 8.95 2.42 3.22	4.76	12.30 8.37 3.38 7.95 — —	8.00
Mean effective porosity per zone	12.85		8.29		11.80	

5 to 6%. More extensive data revealed the me relationship, as shown by V.I. Danchev d V.V. Ol'kh [4]. We are inclined to regard e inverse relationship between uranium con- nt and effective porosity as the result of a ling of the pores by an ore complex in the ocess of ore formation, although these thors explain this relationship in somewhat fferent fashion.

The physical properties of the organic atter in the ore complex class it with the lid bitumens. This is usually a dark brown black, viscous or hard mass, virtually soluble in organic solvents.

Analysis of the organic matter performed V. A. Uspenskiy shows that the ratio of rbon to hydrogen (C/H) fluctuates in various actions from 7.0 to 15.6 and amounts to 10, the average. This ratio characterizes ghly-polarized hydrocarbons of the petroleum ries, approximating asphalts [5, 17].

Thus, the form in which organic substances e found, and their physical-chemical properties make it possible to regard it as a petroleum derivative occurring under conditions of oxidation.

The organic matter defines the physical location of the uranium mineralization. This is readily seen by comparing Figures 1b and 1f. Increased iron and vanadium contents are observed in the same sections of Subzone IIIB.

In characterizing the ore zone as a whole, the following specific features should be noted.

1. The concentrations of uranium and silicon within its bounds are clearly secondary in nature relative to the enclosing rock.

2. The coincidence in the space occupied by the uranium mineralization and the zone of intensive development of stylolite joints and other solution structures. This circumstance testifies to the fact that an environment favorable to the precipitation of uranium is not favorable to the existence of $CaCO_3$ in the solid phase.

3. Paragenesis of uranium mineralization with oxidized organic matter of the petroliferous series is always found within the limits of the metalliferous zone.

4. It has been found that the elements associated with uranium mineralization are vanadium, cobalt, nickel, and molybdenum. We know that all of these elements are everywhere associated with petroleum and are usually discovered in spectroscopic studies of petroleum ash [8, 24].

Zone IV is a direct continuation of the metalliferous zone. Within it we observe a very gradual transition from commercial to very low-grade ores, and then to uranium contents measurable only in terms of Clarke quantities. The primary lithologic-facies features of the intrusives are retained without change over its entire extent. At the same time, the solution structures, which are so varied in Zone III (stylolitic joints, sutures, microstylolites) are almost entirely lacking within its bounds. The geochemical aspect of rocks of this zone reflects the conditions of a reducing environment that may have existed here for a longer time than when the other zones were formed. The rocks are gray in tone, contain disseminated organic matter, and finely dispersed pyrite. In isolated places one observes spots, intercalations and lenses of secondary bitumens, giving the rocks a brown color.

It is not possible to trace the further development of geochemical zoning as it leaves the ore deposit. We know that commercial petroleum deposits may be found in deeper occurrences of these horizons.

The zoning of the metalliferous level described above is, in general, secondary and epigenetic in nature. The zones described are interrelated by a number of common signs, and there is every basis for considering their origin to be the result of some single process. The presence of organic matter of the petroleum series impels one to believe that there is an intimate relationship between that process and the formation and destruction of petroleum deposits.

It is demonstrated that petroleum deposits out of contact with the surface may be destroyed as a consequence of the intensified circulation of contained water. At the oil-water surfaces, a process of anaerobic oxidation of organic matter occurs, as described in papers by A. S. Uklonskiy [15], L. D. Shturm [21], etc. Infiltration waters containing free oxygen, penetrating from the surface to the depths of the petroliferous strata, expend this oxygen in oxidizing ferrous iron, organic matter, and other components. At greater depths, the oxidation of the organic matter occurs by biochemical means, chiefly due to sulfates contained in the contained waters. As a consequence of the life cycle of desulfurizing bacteria, the oxygen of the sulfates is expended upon oxidation, and the sulfur is reduced and becomes hydrogen sulfide. It is common knowledge that the waters of petroleum deposits contain much H_2S. Under these conditions one may anticipate extensive production of secondary sulfides in oil-bearing rocks.

In the opinion of Uspenskiy and Radchenko [17], anaerobic oxidation causes the petroleum to undergo a gradual conversion to maltha and then to asphalt. As a consequence, the viscosity and specific gravity of the petroleum usually undergoes a pronounced increase in the outer areas of oil pools [10].

The oxidation of organic matter is usually accompanied by the evolution of carbon dioxide which, upon entering contained waters, combines with hydrogen sulfide and organic acids to reduce substantially the pH of these waters [21]. There are also indications that the solubility of calcium carbonate increases sharply in the process of the reduction of sulfates, and that, in this connection pores, cavities, and karst areas develop. On the other hand, one frequently observes, at some distance from the oil-water surface, a precipitation of secondary calcite. Petrographic studies of the carbonate reservoirs of the Volga River country near Kuybyshev have also made it possible to determine the presence of silicon cement in the oil-water surface zones of certain petroleum deposits [1].

In the final analysis, these geochemical processes may result in the formation of the sealed petroleum deposits found widely in the areas of the Second Baku.

We thus arrive at the conclusion that the phenomena of secondary calcitization, solution, silicification, leaching, pyritization, and concentration of the solid bitumens so characteristic of a number of uranium deposits are common in the vicinity of decaying petroleum deposits in carbonate reservoirs.

II. CHANGE IN THE COMPOSITION OF WATERS IN THE DISINTEGRATION OF OIL POOLS

A number of oil pools currently undergoing the destructive action of contained waters are found in the vicinity of uranium deposits and in the same limestone beds. Hydrogeological studies performed on various petroliferous structures make it possible to explain the characteristic features of the contemporary process of anaerobic oxidation of petroliferous bitumens and the changes that result in the composition of infiltration waters.

Table 3

Composition of Contained Waters

Site of sampling	No	Chemical composition of waters (gases mg/lit)	pH
Flowing spring in region of infiltration	1	H_2S — lacking, $M_{0.5}\ \frac{HCO^{3}_{70}SO^{4}_{21}}{Ca_{37}Mg_{34}Na_{28}}$	8.3
Flowing springs near oil pool	2	H_2S — present, $M_{2.4}\ \frac{HCO^{3}_{60}Cl_{39}}{Na_{77}Ca_{12}}$	7.8
	3	$\Sigma H_2S_{123}CO_{110}M_4\ \frac{HCO^{3}_{47}Cl_{40}SO^{4}_{13}}{Na_{88}}$	7.3
	4	$\Sigma H_2S_{40}CO^{2}_{34.5}M_{6.7}\ \frac{Cl_{87}}{Na_{62}Mg_{20}Ca_{19}}$	7.2
Outer zone of water-oil surface	5	$\Sigma H_2S_{87}CO^{2}_{255}M_{12.5}\ \frac{Cl_{86}SO^{4}_{10}}{Na_{87}}$	6.5
	6	$\Sigma H_2S_{63}CO^{2}_{116}M_{17}\ \frac{Cl_{93}}{Na_{87}}$	6.8
	7	H_2S — present, $M_{34}\ \frac{Cl_{94}}{Na_{74}Mg_{14}Ca_{12}}$	6.7
Beyond inner outline of oil pool	8	H_2S — lacking, $M_{84}\ \frac{Cl_{99}}{Na_{94}}$	7.1

The results of the study are presented in ble 3, where the data of analyses of contain- waters are presented in the somewhat dified formulas of Kurlov.

Analysis of Table 3 reveals the following ularities. The total mineralization of the ers increases as one approaches the oil l. As this occurs, the nature of the min- lization changes from magnesium-calcium rocarbonate to sodium chloride with inter- diate waters of complex composition.

Against the background of the gradual change he general mineralization and chemical nposition of the waters, their gas composi- 1 changes more sharply. The oxygen waters he region of infiltration are replaced by rogen sulfide-carbonate waters as one ap- aches accumulations of petroleum hydro- bons. H_2S and CO_2 attain their maximum centration in waters containing petroleum, , in the zone of the water-oil surface where process of biochemical oxidation of petro- n obviously proceeds most intensively. An rease in the concentration of acid gases kes for a reduction in the pH.

A proof of the fact that the reduction of the occurs under the influence of these gases, not of any organic or other inorganic acids tever, is provided by the following experiment. In samples in which an oil film was left on the surface of the water, thus protecting it against oxidation by the air and, in part, against degasification, low pH was retained for half a year. In samples where this film was lacking, the pH increased rapidly to 7 upon aeration.

Beyond the inner boundary of the oil-bearing stratum, where H_2S is completely lacking in the waters, and CO_2 is found in very small amounts, the pH of the waters again rises to 7. Such waters are most highly mineralized.

The principles of change in the composition of waters found within a single petroleum pool have been checked and confirmed from the data of a number of other petroliferous structures. The change in the composition of the waters points everywhere to the reality of the process of anaerobic oxidation of the petroliferous organic matter currently taking place in the same carbonate levels with which uranium deposits are associated.

Study of the chemical composition of contained waters in the process of decay of contemporary petroleum pools will assist in an understanding of many phenomena related to the analogous process in the past, and in recreating some of the features of the environment of ore formation.

III. ORIGIN OF ZONING AND GENESIS OF URANIUM MINERALIZATION

Study of the underground waters of contemporary oil pools to clarify the epigenetic zoning of uranium deposits may lead to the conclusion that the geochemical characteristics of the various zones occurred as the result of the effects on them of contained waters of specific composition.

In all probability, Zone I corresponds to the region in which infiltration waters containing free oxygen have long been at work. The dominance of an environment of oxygen oxidation determined the fact that this zone is enriched in iron hydroxides, and the intensive circulation of water resulted in the formation of pores and cavities in the rock. The lower boundary of Zone I corresponds to complete loss of the free oxygen dissolved in the waters. This is clearly seen in the rocks in the form of replacement of ferric hydrates by sulfides.

Zone II may be regarded as the arena of biochemical oxidation of organic matter past the outer fringe zone of the oil pool.

Most probably, it corresponded to the region rich in waters approximating, in composition, samples 2, 3, and 4 in Table 3, the pH of which fluctuates between 7. 2 and 7. 8. Hydrogen sulfide contamination in the contained waters is reflected in the distribution of the iron sulfides in the rocks. The oxidation and loss of organic matter, i. e., the leaching of the rocks, may be explained by reduction of the SO_4 ion in the waters by the process:

$$SO_4{}^{''-} + 2\,C + H_2O \rightarrow H_2S + 2\,HCO_3{}^-.$$

Is is characteristic that, along with the phenomenon of rocks going into solution (microstylolites, brecciation), we also see secondary calcitization here, reflecting unstable carbonate equilibrium in the waters. The CO_2 generated by oxidation of organic matter induced diminution in the pH of the solution and led to solution of the carbonates in various parts of the stratum. Neutralization of the weakly acid solutions in the carbonate medium facilitated precipitation of the secondary calcite.

Zone III is that in which the ancient oil-water surface was stable. Here, along with the iron sulfides, highly oxidized organic matter is preserved, and silica, vanadium, and uranium are concentrated. The limestones bear the traces of active solution in the form of stylolite joints, sutures, and microstylolites.

The combination of all these signs will serve to explain the hydrogeochemical environment usually observed in the zone of water-oil surface in decaying oil pools.

As may be seen from Table 3, this zone usually contains carbonic acid — hydrogen sulfide waters, the pH of which approximates 6. 4.

The processes of biochemical oxidation of organic matter, proceeding on a large scale, introduced such a large amount of CO_2 and H_2S into the water that the situation that had developed facilitated carbonate loss and interfered with secondary calcitization of the rocks. This same weakly acid medium favored settling of the silica and formation of silica concretions. As is evident from Figure 1b, it is here, too, that the concentration of uranium took place. The chemical aspect of this process will be examined below.

The asymmetrical shape of the ore bodies confirms the fact that the precipitation of uranium and silicon took place against the background of contained waters in motion. The maximum concentrations of uranium and silica, coinciding with the most highly bulged portion of the ore body (Figure 1b, c) are found in the stratum having the maximum effective porosity (Table 2), and the general orientation of the "rolls" within the bounds of this deposit correspond to the probable direction of the movement of the contained waters. Analogous data have been obtained by D. R. Shaw [20].

The concentration of vanadium, nickel, cobalt and other elements in the ore bodies is most readily explained by processes of oxidation and polymerization of petroleum bitumens. We know that these elements are always associated with petroleum bitumens.

Zone IV is apparently a section of a former oil pool. This pool was subsequently destroyed as a result of the tapping of an oil-bearing stratum. Confirmation of this hypothesis is provided by residual oxidized petroleum bitumen in the rocks of this zone.

The dimensions of the zones and the intensity with which epigenetic processes occur may vary in accordance with specific geological conditions. For example, there are cases in which silicification acquires purely mineralogical significance and is manifested in the form of microscopic neogenesis.

The distribution of uranium deposits of similar type over broad territories, and their stratigraphic association with specific levels gives reason to assume the presence of initially disseminated and very low-grade uranium contents, syngenetic with the stratum of carbonate rocks. The motion of the contained waters, the decay of petroleum deposits and the development of lithological and geochemical zoning would necessarily have led to considerable displacements of the elements within the limits of the individual beds.

It has been established that the conditions oxidation and reduction constitute one of the incipal factors governing the behavior of anium in underground waters. It has been own in the work of Germanov and others] that, in the zone of abundant infiltration, iters containing a high positive redox tential, uranium oxidizes to the hexavalent d goes into solution vigorously. On the other nd, in places where oxygen-bearing waters counter accumulations of petroleum, the re-x potential diminishes, and uranium is pre-pitated.

If the behavior of uranium in waters were termined only by redox conditions, the ore dies of the deposit would be close to the ver boundary of zone I (Figure 1e) where iron droxides are replaced by pyrite. Ore bodies arly associated with the oxidation-reduction undary are quite familiar. Vickers [26], for ample, is one who described them.

In our case there is, between zone I and the talliferous zone, a zone II, where a re-ction environment for iron exists, but anium does not yet precipate. The absence a concentration of uranium in this zone is plicable as follows.

The most probable form of transport of anium into a carbonate medium would consist uranyl carbonate complexes. The destruc-n of the complexes with precipitation of npounds of difficult-solubility occurs with alization of the solution to a pH of 11 [13] or h acidification to a pH of 5 or 6 [9]. As was nonstrated above, this reduction in the pH curs within the limits of the water-oil sur-e (zone III). It is here that precipitation of anium from the underground waters occurs.

Thus, the occurrence of uranium minerali-ion against the background of epigenetic ing of a metalliferous level is in agreement h the conditions of destruction of uranyl-bonate complexes in the weakly acidic re-ing environment that occurs at the water-contact in biochemical acidification of roleum hydrocarbons.

* * *

The data in this article testifies to the fact t ore bodies in this type of deposit are epi-etic relative to the enclosing rock. The nium mineralization and accompanying ondary changes occurred as a result of the ect of infiltrating contained waters upon roliferous carbonate rocks. Uranium min-lization occupies a predictable position in established epigenetic zoning of the min-lized stratum. The complex of signs of genetic change in the rocks characteristic of s zoning may be employed as prospecting teria for this type of uranium deposit.

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HE CHARNOCKITES OF BUNGER OASIS (EASTERN ANTARCTICA)[1]

by

M. G. Ravich, and V. G. Kuno

This article is devoted to two distinctive arnockite massifs in Bunger Oasis. They ere discovered by M. G. Ravich and D. S. lov'ev in the process of making a 1, 200, 000 ological survey of the oasis during the ntarctic summer of 1957 [1].

The southern massif, trending northeast-ırd, is exposed over an area of about 15 $ı^2$ (Figure 1) in bipyroxene plagiogneiss. rect contact with the latter can be traced ly in the western portion, since the north d east parts of the massif are bounded by ılts, while on the south it is hidden beneath ayer of ice. Its contacts with the surround-g plagiogneisses are very frequently indis-ct. No quenched borders were ever ob-rved within the endomorph. It is important, wever, that intersecting contacts were seen spots, since they give evidence as to the rusive relationships between the massif and plagiogneisses. No xenoliths of the untry rock were present. Thin veinlets (5 10 cm thick) of aplitic and pegmatoidal ıcogranites were a very rare find.

The northern massif, virtually meridional elongation, comprises the islands north of Bunger oasis, of which the largest are arnockitovyy and Cachalot (Figure). The ıssif is exposed over an area of 30 km^2, the gest portion being concealed beneath a layer ice and the waters of the fjord. Contacts h the country rock were observed only on achalot Island, where the massif abuts on otite-garnet plagiogneisses. The nature of contacts is intersecting, but insufficiently stinct, and no quenched border was found in endomorph. In the exocontacts of the ıssif it was not uncommon to observe zones highly brecciated gneisses and crystalline ıists, for which intensive biotitization and ıphibolization were characteristic, as well enrichment with magnetite. On the western edge of the massif there is a belt of finer-grained basic charnockites, which is trace-able in the submeridional direction for nearly 15 km. Characteristic of this belt in places is a high degree of pulverization of the rocks and an abundance of veins of alaskitic and aplitic granites. Veins, 1 to 6 meters thick, extend in the latitudinal and meridional directions, forming an intricate network, in which the in-tervals are 20-25 meters apart. In the eastern portion of the massif, the veins of leucogranites are rare and pegmatite veins are more frequent. However, fragments of the country genisses are present everywhere and, as a rule, are sur-rounded by biotite-magnetite margins.

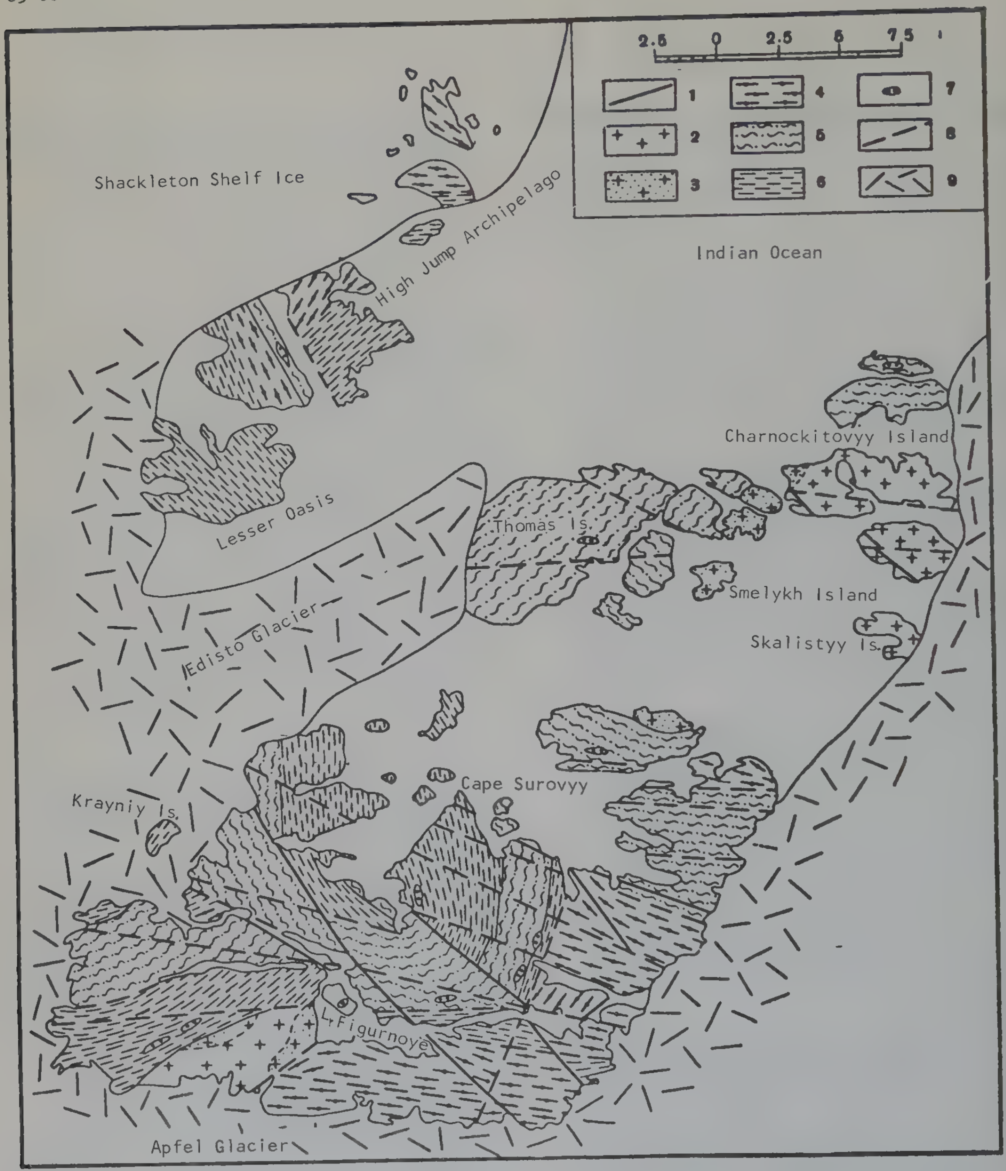
PETROGRAPHIC DESCRIPTION

The charnockite massifs of the Bunger oasis consist primarily of two types of rocks, most closely corresponding to hypersthene norites and hypersthene diorites. These rocks are silicified and feldspathized in varying degrees: the norites to a smaller and the diorites to a greater extent. As a result, it is possible to differentiate subordinate varieties approximating the essexites, the quartzy syenite-diorites, and even granosyenites. Characteristically lacking are typical hypersthene granites, designated charnockites by their initial discoverer, T. Holland [4].

Basic charnockites are represented by massive fine-and medium-grained dark gray rocks with brownish tints. They occur primarily in the peripheral zones of the massifs (Figure 1), where they have undergone comparatively little alteration. Therefore, they have the best preserved, relict, primary, magmatic panidio-morphic-granular and, less frequently, hypidio-morphic granular structures, undergoing transi-tion, in some places, to gabbro-ophites. The maximum degree of idiomorphism is displayed by plagioclase in elongated prismatic plates measuring 2 to 4 by 1. 0 to 1. 5 mm. The grains of rhombic pyroxene, intimately intergrown with monoclinic, are usually of more irregular and sometimes of equidimensional form, and do not exceed 2 mm in size. Porphyry varieties are

[1] Charnokity Oazisa Bangera (Vostochnaya arktida), (pp. 64-77).

65°68'



100°28'
66°21'

FIGURE 1. Geological Sketch Map of Bunger Oasis
Compiled by M.G. Ravich and D.S. Solov'yev, 1959

1 - dolerite dikes; 2 - neutral charnockites; 3 - basic charnockites; 4 - bipyroxene migmatized gneisses (including shadow-weathered migmatites); 5 - migmatized garnet-biotite sillimanite and cordierite gneisses (including shadow-weathered migmatite); 6 - bipyroxene melanokratic plagiogneisses and crystalline shales (feldspathized to varying degrees); 7 - scapolite-diopside rocks, calciphyres and marbles; 8 - faults; 9 - continental and shelf ice. Direction of cross-hatching in symbols corresponds to strike of rocks.

encountered, the structure of which is defined by large prismatic platelets of plagioclase, immersed in a mass of grains of plagioclase and rhombic pyroxene of millimeter size. The phenocrysts of plagioclase are somewhat deformed, have fused edges and, in places, a zonal structure, in which the number of thin zones reaches 15 to 20.

In the rocks described we occasionally encountered recrystallized sections, usually of finer-grained structure, inasmuch as recrystallization apparently preceded cataclasm. The structure of these sections is granoblastic, comprised of xenoblastic grains of rhombic pyroxene and somewhat more isometric grains of plagioclase. Fine quartzy xenoblastic orthoclase aggregates of metasomatic origin are found in these sections. Everywhere we see traces of cataclasm, expressed in the fissures and folding of many plagioclase platelets of elided twinning structure and of spotty extinction. The pyroxene grains are usually highly fissured, and the fissures are normal to the elongation of the mineral.

The composition of the basic charnockites is not constant: 35 to 70% plagioclase (usually 50-55%), 15 to 40% rhombic pyroxene (most frequently in the 30-35% range), 1 to 18% monoclinic pyroxene (usually 3 to 4%), 1 to 6% ore mineral (usually 3 to 4%, 0.1 to 1.5% apatite, 0.5 to 9% biotite (usually 2-3%), 0 to 7% amphibole, 1 to 10% orthoclase (most frequently 3 to 6%), and 2 to 10% quartz (usually 5 to 6%). Flakes of sericite and a fine granular calcite aggregate are found in the plagioclase fissures, while flakes of biotite and ore mineral are found in the pyroxene fissures.

The composition of the plagioclase ranges from No. 55 labradorite to No. 74 bytownite, the most typical being labradorite No. 60. Rhombic pyroxene is represented by bronzite transitional to hypersthene, with a 25-30% iron component (the optical constants of the minerals are presented below in Table 1). In the varieties containing the highest orthoclase and quartz contents, the iron content of rhombic pyroxene increases to 55%. Monoclinic pyroxene is represented by comparatively low-iron diopside in which the hedenbergite molecule represents only 10 to 15%. In feldspathized and silicified varieties this percentage increases to 35%. However, spectral analysis shows that it contains up to 3% aluminum in addition. This makes it possible to classify this mineral simultaneously in the diopside-augite series as well. As a rule, the grains of monoclinic pyroxene are xenomorphic in appearance and form concretions with larger, isometric grains of the rhombic order. Characteristic of many grains of rhombic pyroxene is a graphic intergrowth with monoclinic pyroxene in the form of closely-spaced, thin plates oriented along the cleavage cracks and similar in morphology to the perthitization of plagioclase. Less commonly one encounters spotty inclusions of monoclinic pyroxene in the rhombic. When this occurs it is common for all the intergrowths of the monoclinic in a single large grain of rhombic pyroxene to be of identical optical orientation. When rhombic pyroxene recrystallizes it is common for the intergrowths of the monoclinic gradually to disappear.

It is only in connection with the processes of recrystallization and metosomatism in the norites that amphibole, biotite, orthoclase and quartz develop in norites (Figure 2). The amphibole is represented by common green hornblende of rather constant composition ($Ng = 1.700$, $Np = 1.680$, $2V = -78°$, $cNg = 16\text{-}18°$), which develops most frequently in dissemination among the monoclinic pyroxene, forming complete pseudomorphs both among the grains of the latter and among its intergrowths in the rhombic pyroxene. Biotite is found in any sample of norite, but sometimes only in the form of isolated leaflets, and more

FIGURE 2. Norite

Thin intergrowths of monoclinic pyroxene in the rhombic.
Section 936, crossed Nicol, magnification 22X.

frequently as aggregates in the interstices between the major minerals. As a rule, biotite leaflets are substituted in the edges of a grain of rhombic pyroxene, and are closely associated with ore mineral, the skeletal grains of which are usually fringed with a biotite selvage. In the quartzified varieties it is not uncommon to find symplektitic concretions of biotite and quartz. The biotite is characterized by variable composition expressed in a range of total content of ferruginous material of between 30 and 40%. The orthoclase is unequally distributed in the intervals between the major minerals. Its xenoblastic grains, 1 to 1.5 mm in size, most frequently resorb the edge of the

Table 1

Relationship of Composition of Plagioclases and Colored Minerals to Orthoclase and Quartz Content of Charnockites

Sample No.	Rock	Orthoclase, volume %	Quartz vol. %	Plagioclase no. [1]	Optical constants of the minerals[2]							
					Rhombic pyroxene			Monoclinic pyroxene			Biotite	
					Ng	Nb	Ferruginous comp. mol. %[3]	Ng	Nb	Ferruginous comp. mol. %[3]	Ng	Ferruginous comp. mol. %[3]
752	Basic charnockites	Occasional grains	Occasional grains	74	1,698	1,686	26	1,704	1,681	12	1,628	33
750a		»	2	68	1,705	1,692	32	—	—	—	—	—
750		»	2	64	1,709	1,696	36	—	—	—	1,629	34
446		»	6	60	1,717	1,703	44	1,712	1,688	25	—	—
452		3	7	59	1,718	1,703	45	1,714	1,690	27	—	—
936		6	6	58	1,718	1,704	45	1,716	1,692	30	1,637	42
942		7	5	58	1,716	1,702	43	1,718	1,692	31	—	—
943		8	5	58	1,712	1,698	40	1,712	1,688	25	—	—
217		9	5	55	1,730	1,720	55	—	—	—	—	-
753a	Neutral charnockites	14	13	51	1,735	1,718	58	1,726	1,702	46	1,648	52
6176		17	10	45	1,738	1,721	61	—	—	—	1,653	56
681		15	12	43	1,748	1,731	68	1,730	1,703	55	—	—
683		16	10	40	1,750	1,732	70	1,726	1,704	46	—	—
673		20	16	38	1,748	1,731	68	1,731	1,703	56	1,670	71
371		15	10	40	1,745	1,728	66	1,730	1,702	55	—	—
363		17	15	39	1,750	1,733	70	1,732	1,703	58	—	—
372		22	18	35	1,748	1,731	68	1,730	1,706	55	—	—
357		16	15	38	1,752	1,733	72	—	—	—	—	—
352		18	20	38	1,754	1,735	73	—	—	—	—	—

[1] Plagioclase numbers determined on the Fedorov stage; numbers given represent mean of two or three measurements.

[2] Refractive indices determined with an accuracy of ± 0.002–0.003.

[3] Ferruginous component in colored minerals as determined from the diagrams of V.S. Sobolev and V.Ye. Treger.

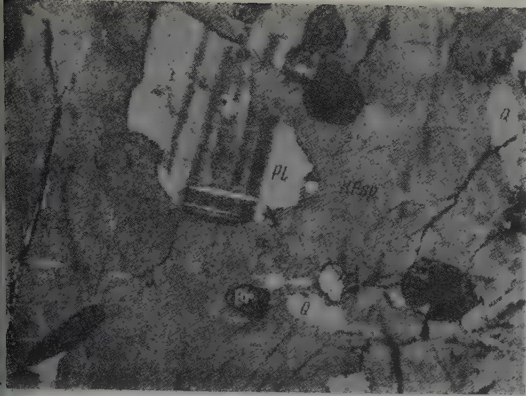


FIGURE 3. Quartzy Gabbro-Syenite Porphyroblastic

Segregations of orthoclase containing resorbed scales of plagioclase, grains of rhombic pyroxene, and quartz. Section 209, crossed Nicol, magnification 54X.

lagioclase leaflets and substitute for them. n some larger orthoclase segregations (Fig-re 3), we encounter plagioclase and rhombic yroxene inclusions or intergrowths of these ninerals. It is not uncommon for flexuous egregations of orthoclase to weld the fissures n the plagioclases. At the interface between he two minerals there is a border of myrmekite uartz intergrowths. In addition to the formation of independent grains, the feldspathization f norites is manifested in the antiperthitization f plagioclase, the leaflets of which are, in laces, rich in spindle-shaped or mottled interrowths of orthoclase. In its optical properties ⊥ (001) Nm = 9-12°, less frequently 14-15°), rthoclase resembles microcline most closely. uartz, associated with microcline and forming mall xenoblastic grains, frequently fills small ssures in other minerals and, more rarely, ccumulates in mono-mineralic lenticles.

The neutral charnockites are represented by oarse-grained, and most frequently porphyryke rocks, brown-gray in color, with a greenh thinge. The more basic varieties comprise the Southern massif; and the more acid, the orthern. These rocks are distinguished by a eteroblastic structure containing residual rismatic granular structure due to prismatic aflets of plagioclase from 3 to 8 mm in ngth and 1 to 3 mm in cross section. These aflets usually have a sinuous outline, since ey have been resorbed by orthoclase segregaons. The porphyry-like appearance of some arieties is due solely to the large prisms of agioclase, and not orthoclase, as in the arnockites of other districts of the Eastern ntarctic. The blastic structures of metaomatic origin are due primarily to the prence of poikiloxenoblastic and usually rather rge segregations (3-5 mm in diameter) of orthoclase and of xenoblastic grains of quartz of various dimensions (diameter from 1 to 3 mm), sometime forming lenticular accumulations. But these rocks also contain granoblastic sections of recrystallized pyroxenes and plagioclases (Figure 4). In these areas grain size is considerably smaller (down to 1 or 2 mm) and glomeroblastic aggregations of colored minerals have been formed. The overall consequence is the formation of rocks combining residual magmatic as well as metamorphic and metasomatic structures complicated by superimposed cataclasis. This latter exists everywhere to a greater or less degree, and is expressed primarily in the fissuring of the rocks. The irregular and frequently branched fissures are filled with oxidized ore minerals and flakes of biotite, sericite and small needles of amphibole. The plagioclase prisms are bent and sometimes broken due to cataclasis. It is characteristic that the new orthoclase formations show considerably milder cataclasis.

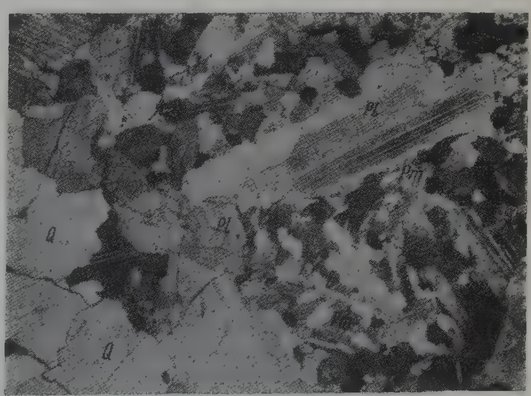


FIGURE 4. Quartz Diorite

Granoblastic areas are enclosed among long prismatic leaflets of plagioclase and xenoblastic quartz grains. Section 693, crossed Nicol, magnification 22X.

The composition of the neutral charnockites is highly varied, primarily due to the superimposed feldspathization, and also due to the substantial recrystallization of the rocks, and is characterized by a 30 to 60% plagioclase content (usually 40-50%), 5 to 20% rhombic pyroxene (most frequently 8 to 13%), 1 to 5% monoclinic pyroxene (usually 2 to 3%), 2 to 6% ore mineral (usually 3-4%), 0.5 to 1.5% apatite, and 1 to 5% biotite (usually 2-3%), 2 to 10% hornblende (most frequently 3-4%), 5 to 45% orthoclase (usually 15 to 25%), and 7 to 25% quartz (usually 10 to 15%). An impurity present in insignificant quantities is iddingsite developed in the rhombic pyroxene, sericite in the feldspars, iron hydroxides and occasional granules of calcite in the rock fissures.

Plagioclase is represented by andesine-labradorite No. 45-52 in the rocks of the Southern massif and andesine No. 35-40 in those of the Northern massif. Highly characteristic is the antiperthitization of the plagioclases, in which the quantity of antiperthite inclusions is from 3 or 5 to 15 or 18% of the volume of the host mineral. Along with the shoestring decomposition antiperthites, one finds, in greater abundance, shapeless or, more rarely, beam-shaped, substitution antiperthites. Not infrequently the orthoclase segregations in the plagioclase are replaced with such intensity that nothing remains but residues, identifiable only by their polysynthetic twins. In these plagioclase grains, myrmekititic quartz intergrowths are as a rule, highly abundant. Rhombic pyroxene falls into the class of ferrohypersthene with a 55-58% ferrous component in the more basic varieties of the described rock group and 68 to 73% in the more acidic. Accordingly, monoclinic pyroxene contains a 42-46% ferrous component in the more basic varieties and 55-60% in the more acidic. It falls into the sahlite and ferrosahlite category, but probably includes the augite molecule, inasmuch as spectral analyses indicate a 2-4% aluminum content therein. The two pyroxenes are closely associated with each other, although the monoclinic pyroxene usually forms intergrowths in the rhombic and, less frequently, is encountered in independent grains. It is characteristic that, the more strongly the rhombic pyroxenes are recrystallized, the fewer intergrowths of monoclinic pyroxene they retain. And, in general, the intensity of recrystallization and feldspathization of the rock will affect the preservation of the pyroxenes, which disappear almost entirely in the most strongly altered rocks. In these it is not rare to find complete pseudomorphs of biotite, hornblende, carbonate, and ore mineral in the pyroxenes.

The ore mineral is represented primarily by titanomagnetite in irregular fine granules, frequently intergrown with pyroxenes or forming a framework around them. Another accessory mineral — apatite — is incorporated in all the other minerals of the rock in fine columnar crystals, although it is more frequently found in large prismatic crystals in pyroxene aggregates. Individual crystalline granules of zircon, surrounded by secondary colored minerals, are encountered sporadically. Biotite is also closely associated with the rhombic pyroxene, and one frequently clearly observes its formation at the expense of the latter, particularly at the boundary between ore mineral and pyroxene. Flakes of biotite outside the aggregations of colored minerals are extremely rare and random. The ferruginosity of the biotite very nearly approximates that of the rhombic pyroxene and, in the case of the more basic varieties of the rock, fluctuates from 48% to 58%, and up to 78% in the more acid varieties. Hornblende is present in xenoblastic granules forming intergrowths with pyroxenes. It becomes pleochroic in the range from the yellow-green and yellow-brown to greenish with a brownish tint. In its optical properties it falls into the category of ordinary hornblende with a total ferruginosity of from 66 to 73% (Ng = 1.696-1.704, Np = 1.670-1.679, 2V = from -72° to -78°; cNg = 16 to 18°). In places, the hornblende is replaced almost entirely by pyroxene, retaining the latter in residual form in its xenoblastic grains. Another generation of amphibole, of bright grass-green color, does not display independent grains, but forms fine borders around the pyroxenes, the ore mineral, and the first generation hornblende. In optical properties it approximates actinolite.

A common orthoclase segregation takes the form of large xenoblastic grains, frequently containing residual plagioclase and rhombic pyroxene and, less frequently, monoclinic pyroxene and ore mineral. It is characteristic that the inclusions are sometimes represented by granoblastic aggregates. This testifies to the formation of orthoclase after the recrystallization of the rock. In some places, vein-like orthoclase segregations weld fissures in plagioclases and pyroxenes. Most of the orthoclase segregations are of microperthitic structure, containing, in places, patterned quartz intergrowths, simultaneously intersecting enclosed residual minerals. The composition of the orthoclase-microperthite is not constant, as is indicated by variations in its optical properties: the extreme 2V values range from -59 to -82°, and the Ng from 1.527 to 1.533. The plagioclase molecule contains not less than 15 to 20%. Microcline is the most widely disseminated representative of, although occasional orthoclase grains are found ($\perp$ (001) Nm = 3-6°). The quartz forms xenoblastic grains of various sizes, irregularly distributed in the rock, and also welds the fissures in the plagioclases and pyroxenes, which it resorbs to a limited degree. Along with the micropegmatitic intergrowths, quartz is found in the form of fine poikiloblastic inclusions.

Xenoliths of biotitized and feldspathized crystallized bipyroxene schists usually form thin bands 1-2 to 10-20 meters long, and the xenoliths of gneiss cataclasites are rounded, measuring from 10-20 to 40-50 cm in diameter. Neither is of defined orientation, and in this respect they differ from the abundant xenolithic bands in the charnockite massifs of other areas of the Eastern Antarctic. Crystalline schists of granoblastic structure consist of labradorite-andesine (30-40%), pyroxenes (30-40%), biotite (5-15%), and orthoclase (2-20%), titanomagnetite (2-6%), and apatite 0.3-0.5%). Hornblende is present sporadically and forms up to 10%, but when this is the case the crystalline schists are wholly unfeldspathized and are

FIGURE 5. Granosyenite

Section 366; crossed Nicol; magnification 56X.

ather weakly biotitized. The gneiss cata-lasites segregate in complex catablastic struc-ures and consist of fragments of plagioclase, nicrocline and quartz, concentrating in longated lenticular sections immersed in a nass of very fine quartz grains, biotite flakes, nd granules of amphibole, carbonate and even oisite, among which one finds fragments of ranite and hypersthene. It is characteristic hat the cataclasite xenoliths are confined to reas where the surrounding charnockites have lso undergone intensive cataclasis.

Veins consist of pink fine and medium granu-ar rocks, chiefly leucocratic. It is char-cteristic that veins are frequently directly djacent to crushed zones in charnockites. hree types of veins may be identified (pro-eeding from the most to the least common):) the finest-grained aplitic granites containing rains measuring 0.2-0.3 mm, comprised al-nost equally of plagioclase, microcline and uartz, with a negligible (not over 2%) content f biotite flakes, and residual grains of hyper-thene and ore mineral; b) the larger-grained laskitoidal granites containing grains mea-uring 1-2 mm, consisting of two-thirds ortho-lase-microperthite and one-third quartz; and) fine-grained pegmatites, formed primarily f prismatic segregations of orthoclase with bundant intergrowths of perthite and graphic uartz intergrowths.

MINERAL PARAGENESIS

The two major types of rocks: noritic and ioritic, comprising the charnockite massifs of ne Bunger oasis, are of identical mineral para-enesis. The rocks themselves, and the bundant varieties thereof, are distinguished nly by differences in the quantitative distribu-ion of the identical minerals, of varying chemical composition. These include plagio-case, rhombic and monoclinic pyroxenes, and biotite. Light is shed upon the genesis of these minerals by comparison of their compositions. They prove to be closely related to the processes of recrystallization and metasomatism (Table 1).

On the basis of the data adduced, it is established that: a) the charnockites unaltered by feldspathization are distinguished by the paragenesis of the basic labradorite with bronzite and diopside augite, i.e., they approximate virtually normal norites and gabbro-norites; b) the weakly feldspathized and silicified charnockites have the paragenesis of andesine-labradorite with hypersthene and sahlite; c) the most widely disseminated charnockites, containing, on the average, about 30% orthoclase and quartz, are characterized by the paragenesis of basic andesine with ferrohypersthene and ferrosilite. In individual varieties, in which orthoclase and quartz constitute about 40%, andesine No. 35 is encountered together with more highly ferruginous pyroxenes. However, when the degree of feldspathization is still more intense, and the charnockites take on a virtually granitic composition (in the vicinity of Mirnyy and Mouson stations), monoclinic pyroxene disappears entirely, and the rhombic is represented by eulite or even by ferrosalite. Biotite, present as an impurity in the basic charnockites as meroxene, in the neutral as lepidomelane, and in the acid as annite, also follows the same law. The content of the ferruginous component in the rhombic pyroxene and the biotite are rather similar.

It may thus be held that feldspathization and silicification were accompanied by a change in the composition of all the other minerals in the rocks and that, in the final analysis, the charnockites, which were basic rocks in their

original form, were converted to neutral and even acid rocks under the influence of caustic-silicic metasomatism.

CHEMISTRY

Ten chemical analyses of the most widely disseminated charnockites of the Bunger Oasis provided a sufficiently distinct picture of the general characteristics of their composition, closely related with the degree of feldspathization and silicification of the rocks (Tables 2, 3, and 4).

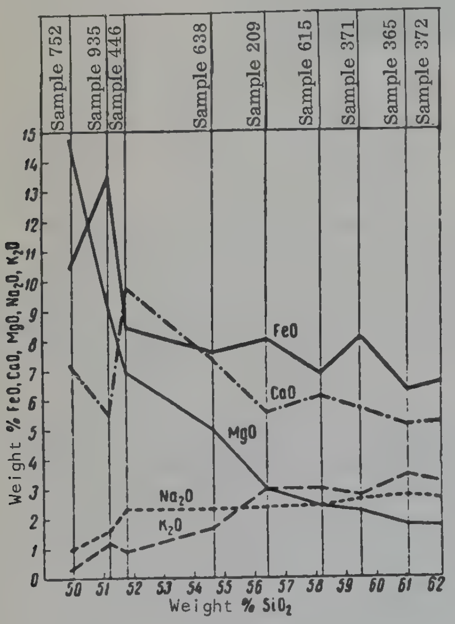


FIGURE 6. Variation in the Chemical Composition of Charnockites

In chemical composition, charnockites have no direct analogs among the magmatic rocks. The major varieties, in terms of caustic feldspars and colored minerals, fluctuate in the range from basic gabbro to gabbrodiorite, but differ from those rocks in that the plagioclases have decreased calcium content, while for the most part they contain free silica, whereas the gabbrotitic rocks are unsaturated with silica. The neutral varieties of charnockites correspond most closely to the diorites and the quartz diorites, but also differ from the latter in the somewhat decreased calcium content in the plagioclases and the increased content of free silica. To summarize, it may be held that all charnockites are silicified rocks, and their feldspar content is not subject to the laws of magmatic crystallization, and is governed solely by the degree of superposed feldspathization. In this connection, we find a more or less constant sodium content to be characteristic of most varieties of charnockites, and a variable potassium content, increasing simultaneously with the increase and content of silica. If we eliminate random variations in the elements relating to the glomeroplasmatic distribution of the colored minerals in the specimens analyzed, diminution of magnesium, calcium, and iron content may be followed quite clearly with increase in silica (Figure 6). This unique behavior of the elements eliminates the possibility of magmatic differentiation in the formation of the charnockites of the Bunger oasis and permits us to see a relationship between its great variety and the silicon-potassium metasomatism of alkaline rocks, under which the addition of silica and potassium and the loss of femic components occurs.

In order to calculate the balance of elements in metasomatism, while eliminating random fluctuations in individual specimens, let us compare the average composition of two groups of charnockites after first recalculating them by the method of T. Barth. The number of cations in a unit cell of basic charnockite (average of three standard analyses) is:

$$Si_{50.5}\,Ti_{1.0}\,Al_{16.7}\,Fe'''_{1.7}\,Fe''_{7.9}\,Mg_{10.2}\,Ca_{7.8}\,Na_{4.1}\,K_{1.5}.$$

The number of cations in a unit cell of neutral charnockite (average of five standard analyses) is:

$$Si_{56.4}\,Ti_{1.1}\,Al_{16.0}\,Fe'''_{1.6}\,Fe''_{5.3}\,Mg_{2.7}\,Ca_{5.6}\,Na_{5.0}\,K_{3.9}.$$

As a consequence of the metasomatism of the rock there is thus to be seen an enrichment by $Si+5.9$ $K+2.4$ and impoverishment by $Mg-7.5$ $Fe''-2.6$ and $Ca-2.2$, with no significant change in the remaining components.

CONCLUSIONS

The charnockite massifs of the Bunger oasis originally consisted of intrusive bodies cutting the country rock with which they now have rather diffused contact. The absence of sharp contacts is explained by the superimposition of metasomatic processes, altering both the intrusions and the country rock, consisting of plagiogneisses of a granulitic metamorphic facies. These intrusions penetrated the country rock after regional metomorphism of the latter, with the result that they contain xenoliths of plagiogneisses that are, however, feldspathized and silicified in a manner similar to that of the charnockites themselves. The charnockite massifs are distinguished by irregularity of composition. One encounters side by side rocks of monzonitic and granosyenitic, syenitodioritic and granodioritic compositions, and only in the endomorphic coating of the massifs, retained along their peripheries, are rocks of

Table 2

Mineralogical Composition of Charnockites

Sample No.	Site where sample was taken	Identification by mineralogical composition alone	Mineralogical composition, volume, %[1]								
			Plagio-clase	Rhom-bic py-roxene	Mono-clinic pyrox.	Ore min-eral	Apat-ite	Horn-blende	Bio-tite	Ortho-clase	Quartz
752	Smelykh Island; western border of northern massif	Basic norite	№ 74 / 55	26 / 38	12 / 2	1	0,1	2	33 / 2	—	—
935	Western shore of Figurnoye Lake; eastern border of southern massif	Quartzy norite	№ 60 / 40	45 / 39	35 / 4	5	1	—	38 / 2	4	5
446	Smelykh Island; western border of northern massif	Quartzy gabbro-norite	№ 60 / 52	44 / 23	25 / 12	2	0,5	—	4,5	1	
638	Northwestern endocontact of southern massif	Quartzy essexitic norite	№ 57 / 62	13	2	3	1	—	3	6	10
209	Central portion of southern massif	Quartzy gabbro-syenite	№ 53 / 48	55 / 11	36 / 1	3	0,5	—	1,5	20	16
615	Southern tip of southern massif	Quartzy dioritic syenite	44	10	1	2	1	3	1	20	18
371	Skalistyy Island, northern massif	» »	№ 40 / 50	66 / 14	55 / 6	3	1	—	1	15	10
372	Charnokitovyy Island, northern massif	Granosyenite	№ 35 / 43	68 / 5	55 / 6	3	1	1	1	22	18
367	» » » »	»	37	6	—	3	1	7	56 / 2	22	21
365	» » » »	Granodiorite	№ 35 / 50	10	1	4	1	2	2	15	15

[1] The numerator states the plagioclase number or the ferruginosity of the colored minerals in molecular %, and the denominator — the mineral content in volumetric %.

[2] Part of the orthoclase is in antiperthitic intergrowths.

Table 3

Chemical Composition of Charnockites

Components	Sample 752	Sample 935	Sample 446	Sample 638	Sample 209	Sample 615	Sample 371	Sample 372	Sample 367	Sample 365
SiO_2	49,91	51.18	51.79	54.70	56.46	58.27	59.65	62.34	62.31	61.19
TiO_2	0.70	1.92	1.16	1.24	2.24	2.04	1.80	1.40	1.38	1.48
Al_2O_3	13.55	11.56	15.62	17.22	15.03	15.36	14.13	13.96	14.10	14.68
Fe_2O_3	0.63	3.37	1.93	2.14	2.68	2,71	2.21	1.77	1.92	2.21
FeO	10.50	13.46	8,42	7.59	8.01	6.90	7.98	6,58	6.63	6.34
MnO	0,28	0,36	0.25	0.23	0,25	0.24	0,23	0.17	0.20	0.19
MgO	14.79	9.10	7.00	5 04	2.97	2.35	2,23	1.81	1.71	1.75
CaO	7.23	5.63	9.79	7.29	5,14	6.06	5.67	5.18	4.80	5.15
Na_2O	0.96	1.63	2.29	2.27	2.42	2.37	2.72	2.72	2.72	2.75
K_2O	0.25	1.15	0.87	1.73	2.96	3.00	2.82	3.32	3.72	3.49
P_2O_5	0.08	0.21	0.16	0.37	0,66	0.75	0.43	0.37	0.36	0.36
Others	1.33	0.89	1.27	1,45	0.78	0.47	0.68	0.93	0,33	0.44
Sum	100.21	100.46	100.55	100.27	100.00	100.52	100.55	100.55	100.18	100.02

Table 4

Comparison of Most Typical Charnockites of Bunger Oasis With "Average" Rocks (Zavaritskiy)

Numerical characteristics	Basic charnockites			Basic gabbro	Neutral gabbro	Acid gabbro (norite)	Gabbro-diorite	Neutral charnockites					Diorite	Quartz diorite
	Sample 935	Sample 446	Sample 638					Sample 371	Sample 672	Sample 372	Sample 367	Sample 365		
a	5.2	6.8	7.6	6	7	8	9	10.1	11.7	10.8	11.3	11.3	11	12
c	4.9	7.7	8.1	8	10	10	9	4.4	4.4	4.0	3.7	4.3	6	4
b	32.1	22.2	19.3	30	25	22	18	16.0	16.0	13.3	13.1	13.2	16	13
s	57.8	63.3	65.0	56	58	60	64	69.5	67.9	71.9	71.9	71.2	67	71
Q	+2.3	+5.9	+6.7	−3	−5	−6	+1	+14.4	+8.0	+18.7	+17.0	+15.5	+6	+14

the gabbro-norite type concentrated. Study of the mineral parageneses of the charnockites establishes that the great variation encountered is due only to a difference in the degree of metasomatic feldspathization and silicification. As metasomatism increases, the content of the ferruginous component increases in pyroxenes, and the basicity of plagioclases diminishes.

The combination in charnockites of residual magmatic and superimposed metamorphic and metasomatic structures testifies to the fact that metasomatism occured under conditions of recrystallization of the rocks, and that the further this recrystallization went, the more intensive was the manifestation of metasomatism. On this basis, and also if we take into consideration the composition of the rocks in the endomorphic halos of the massifs, we may assert that rocks of alkaline composition were the primary sources of the charnockites. On the basis of a study of the chemical composition of the charnockites it may be considered that, in the process of metasomatism the basic rocks were enriched in silicon and potassium and impoverished in femic components. In this connection, we observe the appearance of contact zones of country rocks highly biotitized and enriched in amphibole and ore minerals, as well as biotitic-magnetitic fringes bounding the xenoliths of the plagiogneisses.

The mineral parageneses and metasomatic alterations in charnockite massifs are analogous to those for the surrounding plagiogneisses. It is characteristic that the absolute age of the magmatized plagiogneisses and charnockites of Bunger oasis coincides and is in the 650,000,000-750,000,000 year range, while the plagiogneisses distant from the charnockite massifs and not affected by these processes, are a billion to 1,100,000,000 years old [2]. This testifies not only to coincident conditions of formation but also to the fact that the migmatization of the plagiogneisses and charnockitization of the basic magmatic rocks occurred at the same time. It should be emphasized that the feldspathization and migmatization of the basic cipyroxene plagiogneisses terminates, not infrequently, in the formation of the so-called "shadow-weathered" migmatites widely distributed in the southern part of the Bunger oasis. These shadow-weathered migmatites can also, as a matter of fact, be called charnockites, but of granitic composition, inasmuch as they contain residues of highly ferruginous rhombic pyroxene in the quartzy feldspar mass. Shadow-weathered migmatites everywhere retain the residual eutaxitic texture of plagiogneisses, are represented by fine-grained rocks and, of course, contain no residual magmatic structures, inasmuch as they developed from metamorphic rocks.

In other words, charnockites do not constitute a distinctive group of magmatic or metamorphic rocks, but developed as a result of the ultrametamorphism of basic rocks of the most varied genesis. Charnockitization of basic rocks occurs at considerable depths, under conditions of recrystallization in the solid (and, perhaps in a partially softened) state, with active participation by metasomatic solutions similar in composition to a selective fusion forming the vein material of normal migmatites. Charnockitization must, therefore, be regarded as a special case of granitization.

The country rock around the charnockite massifs did not always undergo intensive migmatization, and even in the massifs themselves the rocks are unevenly feldspathized and silicified. This apparently occurs as a result of the non-uniform degree of cataclasis and the related recrystallization. These processes are an essential prerequisite to metasomatism and govern the intensity with which it proceeds. It is clear that, during differential tectonic movements, the compact massive intrusions are subject to cataclasis to a considerably greater degree than the comparatively more plastic strata of the country gneiss. In the massifs themselves, the abyssal portions are recrystallized to a considerably greater degree than the apical, with the result that the latter have undergone minimal alteration as a rule, and have retained a composition most nearly corresponding to the initial composition of the rocks from which the charnockites developed.

Of course, it is difficult to indicate the sources giving rise to metasomatic solutions in abundance sufficient to granitize enormous thicknesses of rock. However, this same difficulty is inherent in the entire granitization problem. A solution presents itself most readily if the source be considered to be the strata of gneisses themselves (saturated with aqueous solutions to a greater or lesser degree), the most acidic and alumina-enriched varieties of which are capable, under conditions of ultrametamorphism, of yielding colossal volumes of alkaline metasomatic silicon solutions even to the point of selective fusion.

Rocks of the charnockite series are widely distributed in Eastern Antarctica. Their presence on Adelie Land was first reported as far back as 1918 by F. Stillwell, geologist of the Australian Antarctic Expedition of 1911-1914 [5], who classified in this series strata of hypersthene gneisses and bodies of more massive rocks, which he considered to be the products of metamorphic recrystallization under the conditions of the katazone. Tilley [6], who processed the data of the British-Australian-New Zealand Antarctic Expedition of 1929-31 derived from Enderby Land, identified a distinctive type of charnockites lacking in orthoclase, and called them enderbites, believing them to be most probably of primary magmatic

origin. Nockolds [3], who processed the data of the Australian Antarctic Expedition of 1911-1914, published in 1940 a work on the petrography of the rocks from the Queen Mary Land and Wilhelm II Land coasts, in which he classified not only the massive rocks in the vicinity of Mirnyy, but all bipyroxene gneisses and crystalline shales, with charnockites of magmatic origin.

L. V. Klimov, who made a detailed survey of the Mirnyy district in 1957, reports the presence there of a large charnockite massif of complex structure. Individual portions of this massif are distinguished by an exceptional abundance of blocks and fragments (usually in undisturbed bedding) of the country rock reminiscent of agmatites. Such massifs could not have been formed from a normal magmatic melt, but were formed as a consequence of the mobilization of blocks of a metamorphic stratum of granulitic facies most favorable to ultrametamorphism, wherein only partial fusion of the solid substrate occurred. The chief agent of mobilization was the highly mobile melt selectively fused from solid rocks enriched in potassium and silicon. The intrusive relationships between charnockites and the country rocks are made possible by the rather limited mobility of such a "mobilizate". A similar complication of the genesis of charnockites would fall theoretically within the framework of granitization processes.

Thus, the basement of the Antarctic platform, comprised of plagiogneisses and crystalline shales of the granulitic facies of metamorphism, is a region of large-scale distribution of the charnockite series, probably formed in the process of granitization of regionally metamorphosed rocks and the intrusions of alkaline composition incorporated, as is indicated by, to cite an example, the analogous rocks of the Bunger oasis.

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SOME PROBLEMS IN THE MECHANISM OF FORMATION OF THE TUFFS OF THE IRENDYK FORMATIONS[1]

by

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The deposition of pyroclastic material deriver from volcanic activity is of considerable importance in the diverse and complex sedimentary processes in volcanic zones where eruptive activity at times becomes extremely great, scattering large volumes of clastic material over land and sea. Subsequently this material behaves very much like ordinary terrigenous material, being transported and deposited by the same agencies and under similar conditions. It is natural, therefore, that volcanoclastic deposits have much in common with terrigenous, deposits in respect to structure, texture, and stratification. There is, however, an essential difference between the two — the mode of origin of source material, long-term and relatively uniform for terrigenous material as against the short and catastrophic for the volcanoclastic. While the first originates usually in large source provinces, the source of the second is strictly localized. The manner and rate of progress, too, are different. It is only natural that this, too, would be reflected in the structural and textural features of volcanoclastic deposits, and particularly in their stratification.

Our knowledge of the lithology of volcanic-sedimentary formations is still too inadequate to understand all the relationships between the specific features of their volcanic origin and the aspect of their rocks and associations. Youthful continental pyroclastic deposits have been better investigated in that respect but their marine analogs are little known, and many more data are needed to understand the process of eugeosynclinal mechanical sedimentation in its entire scope and diversity.

Of interest in this connection are certain structures peculiar to the Irendyk tuffs and tuffaceous sections which we have studied on the eastern slopes of the South Urals, the Irendyk, Krykta, and Kurkak Ranges, of the Irendyk zone investigated by L. S. Librovich [2].

Some investigators assign the Irendyk formation largely to the Lower Devonian [3]; others, to the Silurian (S. M. Andronov, in a paper read before the Moscow Society of Nature Students), and still others believe that it includes both Upper Silurian and Lower Devonian deposits [4, 5]. Inasmuch as age has nothing to do with the subject of this article, we will say no more on this topic.

This formation is represented by complex combinations of propylitized effusive and volcanoclastic rocks with occasional marbles and siliceous jasper-like formations. The effusive rocks are represented chiefly by augitic, hornblendic, and plagioclase (albite) porphyrites of an andesite- to andesite-dacite primary composition. Basic spilite- and diabase-porphyrites are subordinate. As mentioned by earlier investigators [5], volcanoclastics are predominant over the effusives, on the whole. In some sections (the Irendyk Range at the latitude of the village of Tubinskoye), they make up almost the entire formation; in others, its lower half (Irendyk Range, south of Lake Talkas) or its middle and upper part (Bol'shoy and Malyy Kizil Rivers).

The good exposure of the Irendyk formation affords an unbroken sequence of all of the details of the stratified deposits and the subsequent alteration of the various types of rocks. Especially representative in that respect are the railroad cuts in the Krykty Range, in the Malyy Kizil valley.

Inasmuch as the mechanism and formation conditions of amy clastic sediments are best learned from a study of their structure, texture, and stratification, we shall discuss these features as they relate to the Irendyk formation.

PRINCIPAL STRUCTURAL TYPES OF VOLCANOCLASTIC ROCKS

Decidedly predominant among the volcanoclastics of the Irendyk formation are

[1] Nekotoriye voprosy mekhanizma formirovaniya tufovykh nakopleniy Irendykskoy Svity, (pp. 78-87).

agglomerates and tuffs, largely of an andesite-basalt primary composition (tuffs and agglomerates and tuffs, largely of an andesite-basalt primary composition (tuffs and agglomerates of augite-, hornblende-, and plagioclase porphyrites). There are occasional and peculiar scoria, sometimes cinder tuffs associated with diabasic and spilitic flows. Inasmuch as these are developed in small local layers among the lavas, we omit their description and concentrate only on the andesite-basalt volcanoclastics.

These volcanoclastics, like the effusives, have undergone an intensive greenstone metamorphism; however, their structural features have remained, as a rule, almost intact. In accordance with the predominance of a definite granulometric fraction, we differentiate the following structural types:

Rock	Predominant fragment sizes
Coarse agglomerate	>10 cm
Fine "	1-10 cm
Coarse-grained tuff	1-10 mm
Medium-grained "	0.5-1.0 mm
Fine-grained "	0.1-0.5 mm
Very fine-grained tuff	< 0.1 mm

The agglomerates are green, poorly-sorted rocks consisting of fragments of varying sizes ranging from a fraction of one centimeter to 30-40 cm, variable in form, often irregular, some fragments carrying fine thorn-like projections; the latter suggest a very special means of transportation. The fragment composition is uniform for some layers — pyroxene or plagioclase porphyrites; in others there are in addition diabases and spilites, with occasional tuffs. The cement is a coarse crystallo-lithoclastic tuff. The amount of cement varies from very small to so large that the rock becomes a tuff-agglomerate.

The coarse-grained tuffs are usually lithoclastic, although some layers contain many crystals, with either pyroxenes or plagioclases predominant. These tuffs are very poorly sorted, with fragments ranging in size from 0.2 to 2-3 mm, occasionally attaining 5-7 mm. Despite this wide size range, the extremes are rather rare, with 1-1.2 and 2-3 mm fragments predominant. The lower size limit is the same for both the crystals and lithoclasts, being 0.2-0.3 mm; the upper limit is, of course, different: 0.6-0.8 mm for crystals and 2-3 mm or more, for lithoclasts.

The shape of the fragments is determined by their composition. The lithoclasts are irregular, often contorted in outline, characteristically containing bifurcating spherical cavities looking like the remains of an exploded bubble. Pyroxenes are represented by sharp angular fragments, the plagioclases, by elongated crystals. The tuff cement is scarce and consists of such secondary minerals as chlorite, prehnite, and epidote, apparently developed out of fine ash.

The medium-grained tuffs are similar in composition and shapes of fragments, to the coarse-grained tuffs. They are a mixture of fragments ranging in size from 0.2 to 1.2-1.5 mm, but with the 0.5 to 0.7-0.8 mm sizes predominating. On the whole, these rocks are richer in crystals than the coarse-grained tuffs, although lithoclasts predominate at times. The contrast between their lithoclasts and crystals is not as great as in the coarse-grained tuffs, with both the lower and upper size limits being similar to identical. The cement is also scarce, although more abundant on the whole than in the coarse varieties, and it is of the same mineral composition.

The fine-grained tuffs are rocks in which the 0.1-0.5 mm particles predominate. In general aspect, they are similar to the medium-grained tuffs, except for the fragment size and better soring which is often almost perfect. The grain size in some varieties is 0.1-0.2 mm, 0.1-0.5 mm in the less well-sorted. Their composition differs somewhat from sample to sample — from predominantly crystal fragments (plagioclase and pyroxenes) to strongly altered lithoclasts. Completely chloritized glass fragments are present in some samples. The cement is represented by secondary minerals, scarce to abundant from one variety to another, and often accounting for over 50% of the rock volume.

The fine-grained tuffs (tuffites?) differ from these rocks in external aspect and in composition; strictly speaking, they are assigned to volcanoclastic formations not because of their structural features but because of their constant and regular association with the tuffs and because of their similarity to the latter's cement. Externally, these are blue-green to light-green, homogeneous, fine-grained rocks with a conchoidal to very fine-grained dull fracture. Under the microscope they appear as a fine-grained aggregate of cryptocrystalline silica and chlorite, at times containing epidote, the latter often predominating. This groundmass generally contains small, scattered amounts of very fine (0.1 mm and less) angular grains of highly altered plagioclases and pyroxenes in which occasional radiolaria are found.

Probably, this originally was a very fine ash, now completely devitrified and replaced by secondary minerals. There is no certainty about this, however, since the composition and

structure of sediments have not been preserved. It is quite possible that a very fine terrigenous (argillaceous) material also was present here; in that event, this rock should be assigned to tuffites.

Structurally, i. e., in terms of grain size these tuffs are connected by gradual transitions. In other words, there are a number of granulometrically different rock types whose boundaries are arbitrary. In this series, from the agglomerates to the very fine-grained tuffs, the transition from one type to another is not always the same. Two types of these structural transitions have been observed. The first it characterized by the ratios of coarse- to intermediate- to fine-grained tuffs, each differing in structure from its neighbor only in the predominant size of its fragments. This series is granulometrically unbroken, with the gradation boundaries practically arbitrary. For this reason, it is often difficult to assign a sample to a specific granulometric types.

Transitions in structure are quite different between the agglomerate and the tuff, as well as between fine- and very fine-grained tuffs. Here, they are determined by changes in the quantitative ratios of two, more or less isolated, granulometric fractions, rather than by a gradual change in fragment size. For instance, fragments larger than 1. 5-2. 0 cm almost always predominate in a finely clastic agglomerate, while the coarsest tuff consists, on the whole, of particles smaller than 5 mm. Consequently, there is a clean-cut break between the predominant fractions of these two types. At the same time, the fine-grained agglomerates always contain some coarse tuff (as cement), in some samples in quite small amounts and in others, enough to warrant calling the rock tuffo-agglomerate.

A similar structural transition exists between the fine- and very fine-grained tuffs. The extreme (finest-grained) variety of the first consists for the most part of 0. 1-0. 2 mm particles, although a small amount are about 0. 05 mm. The very fine-grained tuff is formed of pelitic particles. Thus, there is a granulometric break between the principal components of the two types. At the same time, many fine-grained tuffs are characterized by an abundance of cement similar to the very fine-grained tuff groundmass; in some varieties, the amount of this cement reaches 50% and even more. Thus, the gradual transition between these two types is realized not by an unbroken series of grain size gradations but by a quantitative ratio of two sufficiently clearly separated fractions.

These transition differences are probably due to the varying nature of the original eruption products.

The crystal-lithoclastic tuff material arrived at the surface as a mixture of fragments of assorted sizes, but generally varying from 0. 1 to 5 mm. A further differentiation of this material took place in the process of sedimentation.

The separation of the agglomeratic and tuffaceous fractions cannot be explained solely by their differentiation in terms of transportation and deposition. Apparently, it existed at the time of the eruption. Characteristically, the agglomerate fragment often have a diversified composition, with lava-breccias abundant among the latter (Figure 1) and lumps of tuffs occasionally present. The fragments are irregular, angular and the ordinarily typical "volcanic bombs" are not in evidence.

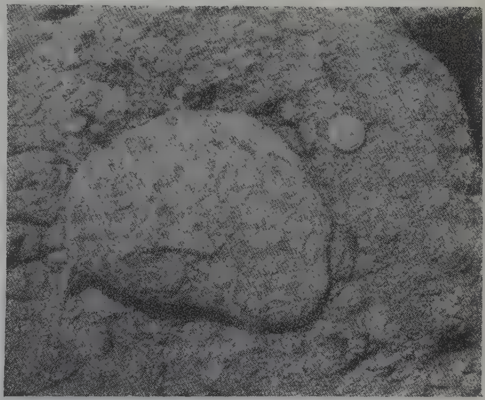

FIGURE 1. A large fragment of lava-breccia in finely clastic agglomerate. Malyy Kizil River.

All this would seem to suggest that the agglomerate fragments originated largely in explosions of previously erupted rocks filling the craters and comprising the volcanic structures (the accessory component of C. Wentworth and H. Williams [10]. The granulometric differentiation of agglomerates and the bulk of tuff fragments can be explained by this difference in origin. It is probable that fragments which originated in the shattering of a volcanic plug and crater walls are present in the tuff fraction; but true pyroclastic material is missing in the agglomerate fraction, at least in appreciable amounts.

The "independence" of the very fine-grained tuff fraction, too, appears to be related to the originally heterogeneous nature of the material. The coarse-, medium-, and fine-grained tuffs consist of lithoclasts and crystal fragments, the lower limit of grain size occasionally reaching 0. 05 mm, but they are usually larger. The very fine-grained tuffs consist mainly of devitrified ash material originally consisting of

glass particles smaller than the litho- and crystalloclastic components (volcanic dust).

STRATIFICATION OF TUFFS

The Irendyk volcanoclastics form thick sequences of thickly-, unevenly-, to fairly evenly stratified series, in which the stratification is asymmetrically rhythmic. This stratification effect is brought about by an alternation of regularly constructed pockets, or "multilayers" according to N. B. Vassoyevich's terminology, often referred to as "rhythms". Each of these multilayers is characterized by a progressively finer grain of rocks, from the base to the top. Two types of multilayers are differentiated.

A complete multilayer of the first type consists of agglomerates, tuffs, and very fine-grained tuffs, with a distinct transitional tuff-agglomerate layer between the first and the second members. A typical multilayer has the following structure (reading upward):

1. Agglomerate, fairly coarse in some bases, in others — less coarse, but always containing a small amount of cement. Its thickness is usually 3-5 m.

2. Finely clastic agglomerate, with abundant tuffaceous cement, whose amount increases gradually but rapidly, upward, until the rock becomes tuffo-agglomerate. The lower contact of this member is fairly sharp (Figure 2); the upper contact is transitional. This member is 2-3 m thick, or less.

3. Massive tuff, coarse-grained at the base, changing to medium-grained in the middle, and fine-grained higher up. Thus, part of this multilayer shows distinctly sorted banding or graded bedding. Its thickness varies in different multilayers from 0.3-0.5 m (seldom) to 3-5 m, and usually is 1.5-2 m.

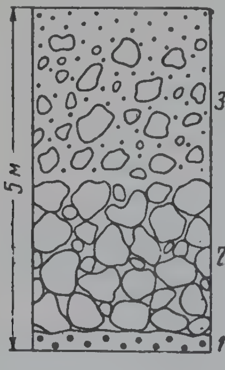

FIGURE 2. Structure of the lower part of a multiple-bed first type polystratum (diagram)

1 - tuff of the underlying polystratum; 2 - coarsely clastic agglomerate; 3 - finely clastic agglomerate with abundant tuffaceous cement.

4. Very fine-grained tuff, stratified, often platy; the stratification is associated with the presence in the fine-grained rock of thin intercalated bands, 0.5-1.0 mm to 0.5-1 cm thick. Characteristically, even these fine intercalations show a distinct graded bedding (Figure 3). As a rule, the stratification is horizontal, but rarely, is cross-bedded or wavy. Some layers have been disrupted by syngenetic deformations — fairly flat in some places, only slightly disturbing the original wavy stratification (Figure 4), to quite sharp in other places where they occur in a system of larger isoclinal plications cut by small faults and thrusts. The very fine-grained tuffs are usually 1-2 to 20-30 cm thick, seldom 0.5-1.0 m. Their lower contact is locally indistinct, since they are associated with a gradual transition (of the second type) to the underlying fine-grained tuffs; in other places it is quite sharp, flat to wavy. The upper contact is always very sharp, since the agglomerate of the next multilayer is in direct contact with the fine-grained rocks.

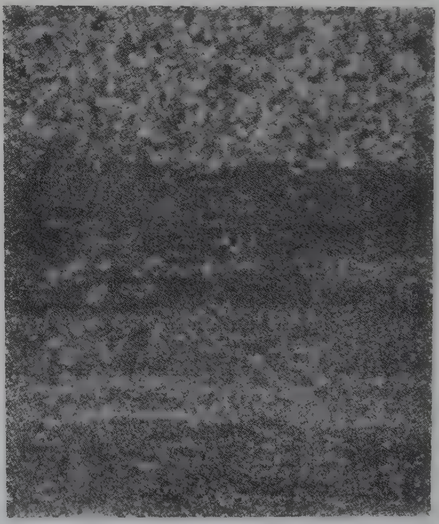

FIGURE 3. Graded bedding in the upper part of a tuffaceous multilayer

One Nicol, X12.

It should be noted that no two multilayers are identical in thickness and structure; the agglomerate is often represented by a thin layer or else is altogether missing as an individual member, and the multilayer begins with a coarse-grained tuff containing dispersed coarse

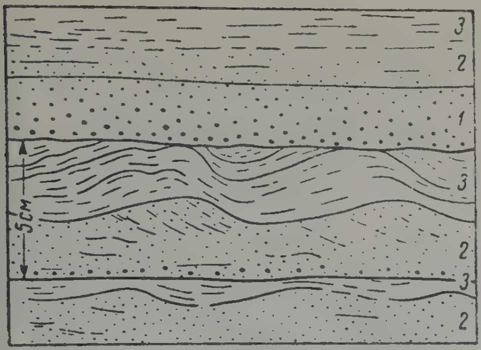

FIGURE 4. Alternation of tuffs with different grain sizes

Diagrammatic sketch of an exposure on the western precipice of Mt. Kartash-Bao. 1 - coarse- to medium-grained tuff; 2 - fine-grained tuff; 3 - very fine-grained tuff with gentle syngenetic deformations.

porphyrite fragments; the very fine-grained member is often missing, and the multilayer terminates in fine- to medium-grained tuff.

of two members and, as we shall see, they are also characterized by certain special structural features.

A second-type multilayer consists of the following rocks:

1. Tuff, coarse- to medium-grained below, changing upward to fine-grained. Small, angular to flat and rounded fragments of fine-grained tuffs often occur at the base. As a rule, they occur in haphazard agglomerations; when present in large numbers, they form a peculiar breccia or conglomerate layer at the base of the multilayer (Figure 5). Characteristically, lenses and intercalations of very fine tuffs, 1-10 cm thick, occur in the coarse tuff, with sharp and often uneven contacts (Figure 6). These uneven contacts are caused not so much by erosion as by a creep of the sediments and by the warping of fine plastic ooze layers under the weight of the coarse tuffs often imbedded in the fine sediment. It is to be remembered that such contacts are common in flysch, and have been called load casts. These intercalations change at times to a finely-clastic breccia (Figure 7) whose origin is best explained

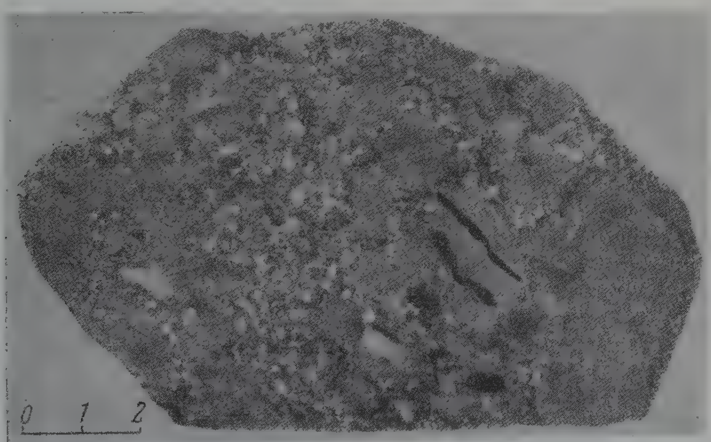

FIGURE 5. Coarse-grained tuff at the base of a multilayer of the second type, containing numerous fragments of very fine-grained tuff; Mt. Kartash-Bao.

by the crumbling of a very fine-grained layer, perhaps already covered by granular deposits during an earthquake.

Although the lower contact of a multilayer is very sharp, as a rule, there are instances of an indistinct contact between the medium- to fine-grained tuffs terminating the multilayer and the overlying agglomerates. The presence of many large fragments in the tuff tend to obscure the contact.

The multilayer of the second type differ from those of the first largely by the absence of agglomerate. As a consequence, they consist

The thickness of this first member varies from one multilayer to another, ranging from 0. 5-1. 0 to 3-4 m.

2. Very fine-grained, micro-stratified tuffs, 0. 2-0. 5 m to 3-4 m thick. Where this member is not eroded, it constitutes an

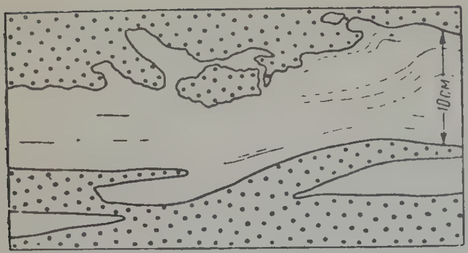

FIGURE 6. Intercalation of very fine-grained tuff (tuffite) in coarse-grained tuff.
Sketch of an exposure along the Bol'shoy Kizil River.

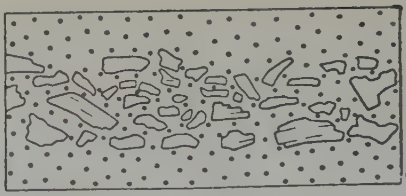

FIGURE 7. Coarse-grained tuff with an intercalation of breccia of very fine tuff fragments.
Sketch of an exposure along the Bol'shoy Kizil River.

essential component of this multilayer in contrast to those of the first type where it is often missing because of non-deposition.

Those sections where the first-type multilayers predominate are, on the whole, coarser and their rhythm is often broken by the appearance of thick (20-30 m) pockets of massive agglomerates or tuffs, with local porphyrite layers. The porphyrites overlie the agglomerates, usually, but not always, and are overlain in turn by tuffo-agglomerates or coarse-grained tuffs.

Sections with predominantly second-type polystrata are comparatively more evenly stratified, although the multilayers of the same section vary greatly in thickness and structure. Here the stratification rhythm is interrupted by platy, thin-bedded patches (20-30 m thick) or by thick beds of massive tuffs.

SEDIMENTATION CONDITIONS OF THE IRENDYK VOLCANOCLASTIC STRATA

Volcanoclastics are especially well developed in the Irendyk formation where they account for the bulk of the sediments. Therefore, the sedimentary conditions of Irendyk time are best understood from a study of the structural and textural features of these deposits.

The essentially marine origin of the Irendyk deposits has long since been established and at present this question is not considered doubtful. [2]. At the same time, a study of the structure of these tuffs sheds light on certain important aspects of their sedimentation, which have been ignored previously. This is probably because the importance of turbidity currents in transporting coarse-grained sediments has become apparent only recently [7]. A detailed description of the coarse deep-water deposits associated with these currents offers a new solution to many paleogeographic problems.

The main sedimentary process in the Irendyk marine basin was the deposition of volcanoclastic material, mostly andesite-basalt in composition. Thus, the principal source of sediments was endogenic. The influx into the basin of huge masses of pyroclastics, exceeding the volume of flows, suggests a predominance of explosive eruptions. The latter are associated with land volcanoes of those located at shallow ocean depths where the hydrostatic pressure is low [1]. It may be assumed, then, that the Irendyk volcanoes, or at least their summits, were located in shallow waters and even above sea level, forming volcanic islands. The volcano sites are marked by a development of thick and coarse agglomerates and an abundance of lavas and lava-breccias. However, as pointed out above, almost the entire Irendyk formation is represented locally by stratified tuffs. In the vicinity of the volcanic craters, these tuffs contain a large amount of coarse agglomerates which become scarcer farther away. The amount of very fine-grained rocks increases in the same direction; these originally consisted of "volcanic dust" with some fine terrigenous suspension, always present in a marine basin, and setting slowly out of the marine water load. Their regular stratification and the pelitic composition suggest a deposition under quiescent hydrodynamic conditions associated with the bottom below the "ooze line".

This unhurried and monotonous sedimentary process was disturbed from time to time by the appearance of coarser volcanoclastic sediments.

It has been demonstrated above that tuffs are thick layers of unsorted coarse material with a distinct, graded bedding. As demonstrated by the work of F. Kuenen [7] and other geologists and oceanographers, these are typical deposits of turbidity currents originating on clearly expressed submarine slopes, on whose upper sections large bodies of unconsolidated sediments piled up for one reason or another. These turbidity currents, unlike other marine currents, are capable of transporting large fragments for long distances, in suspension rather than by saltation as in rivers. In this process, the fragments are not rounded but may preserve

their sharp edges and even the branching protrusions. As pointed out before, such fragments are fairly common in the Irendyk agglomerates and tuffs.

F. Kuenen [9] states that under present conditions, turbidity currents have not been observed above the bathial environment.

One of their features of turbidity currents is their rapid appearance and disappearance — a result of the sporadic nature of their causes. Not the least among the latter are earthquakes, as amply proved by present day observations [8]. Undoubtedly, this was the main factor in the Irendyk volcanic zone. In addition, unlike terrigenous sediments, the inflow of pyroclastic material is sporadic, thus promoting the development of turbidity sedimentation. The deposition of flysch sandstones is a typical instance of the latter; some western European geologists call them "turbidites", to emphasize their origin. Many of the specific features of "turbidites" are well expressed in the Irendyk tuffs: gradation bedding, a regular horizontal stratification, load casts, and creep deformations at the top of the multilayers. At the same time, such typical features of flysch sandstones as ripple marks are missing in the tuffs. The latter are considerably thicker than the flysch sandstones, as a rule, while the intervening pelitomorphic rocks are much thinner.

However, even the flysch formations themselves are known to contain sandstones appreciably different from the typical. In this country they came to be known as coarse flysch and certain Polish geologists proposed the name of "fluxoturbidites" [6]. They are coarser-grained, less argillaceous, and their intervening shales are thinner, if present at all. The appearance of such sandstones is associated with the sudden arrival of an unusually large volume of clastic material or with a sudden steepening of the slope. The authors of the term, "fluxoturbidite", believe that this rock type originates in a peculiar flow — transitional between a turbulent turbidity current and a submarine slide. It appears that the origin of the Irendyk tuff turbidites was often similar to that of "fluxoturbidites".

The question arises as to whether the graded bedding in tuffs is simply a result of the different rate of settling for volcanoclastic fragments of different weights, rather than of the deposition of pyroclastic material by turbidity currents. We believe that such a process is operative in isolated instances but it cannot explain the accumulation of the entire body of tuffs. This is contradicted by the extremely poor sorting of tuff material and by the long distances over which the large fragments have been transported.

A study of the structure and stratification of the Irendyk section has revealed an extensive development of sediments deposited by turbidity currents, i. e., deep water sediments. At the same time, the prevailing explosive nature of the eruptions is an indication of shallow depths. These two facts suggest a strongly differentiated relief of the Irendyk sea bottom — that of a typical geosynclinal basin with volcanic archipelagos and fairly deep troughs.

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ON THE PROBLEM OF THE ORIGIN OF THE RZHEGAKIA (RZEHAKIA) BEDS[1]

by

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The last eight to ten years have witnessed e publication of many new data on the origin d age of Rzhegakia (Oncophora) beds. These iblications contain many contradictions, mostly their treatment of the formation conditions of ese beds.

Such differences in opinion are quite natural cause the Rzehakia beds cover a large area rom Switzerland to the Aral Sea); they have en known for only the past few decades, and e areas covered by them are considerable stances apart (often measured in several ousand kilometers) and are marked by differ- t geological development, which complicates eir correlation.

We attempted in 1954 to correlate the mate- al on these beds and to determine the direction r further studies.

The Rzehakia beds of Switzerland, Bavaria, istria, Moravia, Galicia, and the North Cau- sus are foredeep deposits; those of the Pan- nian basin and the Kura and Rion troughs — e deposits of intermontane troughs.

The molasse province of South Germany is a pical foredeep; it started to develop in the pper Paleogene. This trough was filled with pper Paleogene to Lower Miocene marine, ackish-, and fresh water sediments, and ifted gradually to the north toward the plat- rm. In the process, Tertiary deposits adually retreated northward away from the sing Alpine range. The older deposits are rtly involved in the orogenic movements and e folded; accordingly, two zones are present the Bavarian molasse: the folded and the non- lded.

The Rzehakia beds occur, for the most part, the unfolded outer zone of the Bavarian molasse. They are underlain by thick (up to 2000 m) Bourdigalian and Helvetian marine deposits represented by various marine facies (Upper Marine Molasse). The top of the Upper Marine Molasse includes brackish-water deposits with the Rzehakia, and are assigned to Upper Helvetian by Germany and Austrian geologists [28-30, 34]. The Lower (Upper?) Marine Molasse and the Rzehakia beds are overlain by thick Tortonian-Sarmatian fresh-water, fluvial and lacustrine deposits (Upper Fresh-Water Molasse).

The Rzehakia beds are exposed largely in the north part of the foredeep. Isolated outcrops are found in Rheat-Messkirch province; in the vicinity of Ulm (where they were first described as Kircheberg beds); in the Passau area; and in the southeastern part of the Bavarian plain, at Misbach. A direct connection between the Rzehakia beds in these areas has not been demonstrated and remains hypothetical [32].

In East Bavaria, the Rzehakia beds occur in the upper part of the Upper Marine Molasse represented here by streaky (schlieren) facies whose microfauna suggests a decrease in salinity in the upper intervals. The contact between the schlieren facies and the Rzehakia beds is marked locally by sandstones containing oysters and balanids in some localities. This horizon is only a few meters thick [27].

In the northwestern part of the trough, the Rzehakia beds lie transgressively on Jurassic limestones or on river sands (Graupensand), assigned by German geologists to the Upper Helvetian (Canton Schaffhausen, Switzerland; the vicinity of Auselfingel, at Emmingen, Neudorf, Unterschwandorf, and Hendorf-bei-Messkirch).

The macrofauna of Rzehakia beds is represented by Rzehakia partschi (apparently greatly differing, morphologically, from the Moravian R. (socialis), and representatives of a brakish to fresh-water fauna of Cardium, Cyrena, Congeria, Unia, Bythinia, Hydrobia, Cingula, Neritina, Melania, Melanopsis, etc. F. Sandberger [35] remarked on the faunal changes

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[1]K voprosu o proiskhozhdenii Rzhegakievykh oev, (pp. 88-99).

suggesting a gradual freshening of the basin, and he differentiated the Kirchberg beds into three members: the Cardium, Dreissenia (Congeria — V. Ch), and Hydrobia.

The Rzehakia beds of South Bavaria were studied and described in detail by D. Wittman, in 1957; he differentiated three horizons.

The first horizon, consisting of fine-grained sands with rzehakias and cardids, is 4-45 m thick, and rests on the Marine Molasse. A siltstone marker bed containing the same fauna separates the first and the second horizons. It is overlain by the second horizon of micaceous sands, 15-25 m thick, also with rzehakias and cardids. Its top, too, is marked by a horizon showing evidence of a further freshening; rzehakias and cardids occur here as the genera Congeria, Melanopsis, Neritina, Unio, Planorbis, Ancylus, and even Helix. Above that, there is a siltstone bed (Schillsande) 5-15 m thick, containing the genera Cardium, Congeria, Melanopsis, Neritina, and Unio. Rzhehakias are quite rare, occurring only at the base. These deposits are overlain by sands and marls of the Fresh-Water Molasse, with a sandstone layer carrying unionids at its base.

D. Wittman notes that the marine sandy marls in some exposures gradually change to Rzehakia beds, while the boundary between the two is more distinct in the shallower marine facies of coarse-grained sands. In any event, the appearance of a Rzehakia biocenosis is quite sudden.

The total thickness of the two horizons of Rzehakia beds is about 40 m.

In Lower Bavaria, drill cores contain a sparse foraminifera fauna [29]. Present along with stunted relict species from the underlying Helvetian marine deposits (Globigerina bulloides d'Ord., Nonion scapha Ficht. et Moll., Bulimina elongata d'Orb.) are large Rotalia beccarii (L.) and rare ostracods. The presence of plankton stenohaline foraminifera suggests the presence of the sea. The Rzehakia beds here are inconsistent in thickness — from a few to tens of meters.

In the Molasse zone of Upper Austria, the Rzehakia beds are similar to those in southeastern Bavaria, but here they are as much as 80 m thick.

In South Germany and Upper Austria, the Rzehakia beds appeared at a certain stage in evolution of the Alpine foredeep during the formation of the thick Oligocene-Miocene marine, brackish, and fresh-water sequence. During the orogenies, the foredeep axis shifted northward, the older beds were involved in the tectonic movements and were folded, the folds often being overturned to the north.

A normal salinity sea existed in the foredeep during the Bourdigalian and Helvetian; toward the end of the latter, as a result of orogenic movements, a portion of the Alpine foredeep gradually lost its connection with the sea and became a large embayment open to the southeast. Because of the influx of fresh water carried to the embayment by rivers flowing from the platform in the north and from the young Alpine slopes in the south, the embayment became less saline and was gradually divided into a system of relict brackish water lagoons and lakes into which the Rzehakia beds were deposited.

The transgressive deposition of the Rzehakia beds on older Miocene rocks and even directly on Jurassic limestones and river sands, i. e., on the platform, can be explained by contemporaneous orogenic movements involving a subsidence in the platform margin and a shift of the lagoons toward the platform. This is the so-called "discordant displacement", a phenomenon typical of the evolution of a foredeep [15, p. 15]. At the last formation stage of the foredeep (Tortonian and Sarmatian), the embayment was a plain with a number of flowing lakes and a system of rivers whose deposits completely covered the former trough. At the same time, the orogenic movements weakened and died down.

Much more complex is the problem of the origin and stratigraphic position of the Rzehakia beds in the Alpine-Carpathian foredeep and in Austria and Czechoslovakia. Although the recent works of Austrian and Czech geologists present many valuable data, definite and adequately substantiated conclusions as to the evolution of the basin are difficult to draw. The tectonic development of this province was quite complicated because of its position at a junction of the Alpine and Carpathian foredeeps further complicated by an anticlinal uplift and comparatively young normal faults which brought about the formation of the Vienna basin [18]. Moreover, the marine conditions did not cease here after the deposition of Rzehakia beds, as was the case in South Germany, but persisted as a result of new transgressions. For these reasons, and despite the new studies and drilling carried out here, the correlation of certain Lower- and Middle Miocene beds is often problematical. For the same reason, opinions on the stratigraphic position and correlation of certain Miocene beds have changed repeatedly and radically in recent years. Withal, some data obtained by Austrian and Czechoslovakian geologists bear on the origin of Rzehakia beds. Thus, they have fully corroborated our earlier view that rzehakias in the Grunda beds (Windpassing) are a foreign element, evidently redeposited.

Following a reclassification of the mollusk and foraminifer assemblage in the "Brnen sands",

with rzehakias occurring in a Helevtian marine steno-haline fauna, J. Paulik, J. Tejkal, and I. Cicha [20, 21, 33] have established its Tortonian age. These rzehakias, together with marl fragments carrying a Helvetian fauna, were broken off from the shore, rounded, and redeposited in the Tortonian sea transgression.

Thus, it can be stated more or less definitely that the Rzehakia beds of Central Europe represent an exclusively brackish-water facies; those occurring together with the steno-haline marine fauna are undoubtedly redeposited.

We disregard certain rare marine mollusks which tolerate some reduction in salinity and occur as relicts in Rzehakia beds. This position is shared by such Soviet geologists as L. Sh. Davitashvili [1], R. L. Marklin [11-14], and G. A. Kvaliashvili [4-6].

In Austria, beds with autochtonous rzehakias rest on the "Hall Schlieren" marls, supposed to be Bourdigalian in age. These are brown argillaceous-micaceous sands containing the Rzehakia ([R?] socialis Rzeh.). No microfauna has been observed.

In Moravia, J. Paulik and J. Tejkal [33] relate the Rzehakia beds containing an autochtonous Rzehakia fauna only to those arenaceous deposits in the vicinity of Moravskiy Krumlov which are represented mainly by fine-grained micaceous calcareous sands and sandstones. Collected here, along with Rzehakia socialis Rzeh. and Cardium moravicum Rzeh., were shells of Melanopsis and Congeria; foraminifera redeposited from older horizons; and rare ostracods of the families Cytheridae and Candonidae.

Deposited at the base of the Rzehakia beds are clays, sandy shales, and sands containing fresh- to slightly brackish water fauna of Neritina, Planorbis, Congeria, and Anodonta. In the drill cores of beds under the brackish-water beds (Nosislav Nos. 1 and 2) were marine calcareous clays containing a benthonic microfauna of the "Ottnang" type, assigned to the Lower Helvetian.

Czechoslovakian [21] as well as Austrian geologists [28] regard the brackish-water clays and sands in the vicinity of Znoymo (Czechoslovakia) and the brackish- to fresh-water Zellendorf claystones in Austria, as deep-water equivalents of the Rzehakia beds.

In Czechoslovakia, marly clays and marls containing an Upper Helvetian microfauna occur at the top of the Rzehakia beds, supposed to be Lower Helvetian; the lower interval of Grunda beds occupies this position in Austria. The presence of Rzehakia beds in the interior Vienna basin has not been demonstrated but it is possible, according to A. Pappa.

Czechoslovakian geologists believe that the Lower Helvetian sea transgressed over a limited area of the foredeep. Because of the rapid freshening, the deposits are represented here by various brackish-water facies, and the sedimentation culminated in brackish Rzehakia beds. The sea transgressed once more in the Upper Helvetian, depositing marly clays and marls with a steno-haline fauna. No coarse clastic deposits have been observed at the base of the Upper Helvetian, and the transgression can be inferred only from the microfauna change [19, 20].

It is of interest that here, as in South Germany, Rzehakia beds occur largely on the outer margin of the foredeep, i.e., at the side toward the platform, and seem to have originated in the underlying marine deposits. Their transgressive position on older rocks can also be explained by tectonic movements accompanied by an associated extension of the basin toward the platform. However, we believe that their relation to the overlying marine deposits is still obscure.

Rzehakia beds also occur in another locality of the Carpathian foredeep, within the U.S.S.R. between the lower courses of the Strypa and Zolotaya Lipa, in the vicinity of Galich and Buchach [3, 7-10, 26, 31].

L. N. Kudrin writes [10] that in the Helvetian, isolated areas of the platform and of the outer foredeep zone were invaded by a transgressive sea. This basin changed its outlines throughout the Helvetian and finally separated into brachish-water lagoons and lakes. Deposits of this ingression were represented by thin arenaceous beds containing rzehakias and a marine steno-haline fauna, as well as by fresh-water limestones alternating with calcareous clays. L. N. Kudrin believes that the stratigraphic equivalent of the "Rzehakia beds" is represented by the Steonitz series from the inner zone of the Carpathian foredeep [7, p. 61].

In another work [8] L. N. Kudrin writes that the abundance and variety of the steno-haline fauna in the Rzehakia beds suggest their development in a basin of normal salinity well aerated and lighted, and with enough nourishment. According to L. N. Kudryavtsev, this biocenosis is quite similar to that known from the Rzehakia beds of the Vienna basin and Moravia. This suggests a direct connection between the foredeep and the Vienna basin.

It should be noted that the mollusk association cited by L. N. Kudryavtsev for the Galician Rzehakia beds contains genera from most diverse biotopes (for instance, Chlamys, Congeria, pholatidae). On the whole, the steno-halite faunal elements have a Tortonian aspect.

Very similar depositional conditions are

associated with the Rzehakia beds of the South Ukraine described by I. A. Lepikash [10-a] and more recently by M. F. Nosovskiy [16]. As in Galicia, the rzehakias were found here in thin deposits (0.5-5 m), in association with congerias and a marine steno-haline fauna. Rzehakias disappear completely in the upper parts of the deposit near the village of Kamenka. From their faunal assemblage, M. F. Nosovskiy correlates these deposits with the similar deposits of Galicia and with Grunda beds of Austria. On the other hand, B. P. Zhizhchenko does not rule out the possibility of the rzehakias being redeposited in the Tortonian after erosion of the Rzehakia beds. Since the beds containing the marine steno-haline fauna and redeposited rzehakias have been identified as Tortonian in Austria and Czechoslovakia (Brno sands and the upper Grunda beds), and also because the Rzehakia beds containing the autochtonous rzehakias are an exclusively brackish-water facies, the correlation proposed by L. N. Kudrin and M. F. Nosovskiy is without a basis.

R. L. Merklin proposes [13] that the Rzehakia beds of Galicia are a possible equivalent of the Baranovsk beds and Tarkhan horizon, which are better assigned to the Helvetian.

These differences of opinion show that the Rzehakia beds of Galicia and the South Ukraine warrant a comprehensive study. It is important that their exact stratigraphic position relative to the Tarkhan horizon be determined, with the alternatives of a synchronous or asynchronous redeposition being considered.

Rzehakia beds of South Slovakia and Hungary, located in the northern margin of an interior trough between the Carpathians and Dinaric Alps, had a somewhat different development.

Up to now, they have been observed in only a few places along the north side of the trough: in the vicinity of Modryy Kamen', South Slovakia, and at Shalhotarjan, in northern Hungary. No rzehakias have been found to the south in the Dinaric Alps foothills.

In South Slovakia and northern Hungary, the Miocene starts either with Bourdigalian deposits resting on the Paleogene, or with an obvious unconformity directly on ancient rocks. The Bourdigalian stage is represented by two horizons.

The lower horizons consists of shallow marine sandstones, usually glauconitic; sands, and green-gray tuffaceous clay with intercalations of rhyodacite tuffites and a typical, locally very rich, fauna of large pectinids (Pecten hornensis Dep. -Rom., Chlamys holgeri Geinitz, C. gigas Schloth., C. palmatus Dom., etc.). These are the so-called "Lower Pecten beds" of Hungarian geologists.

Typical of the marine Bourdigalian deposits is the presence of dispersed components of acid effusive rocks (volcanic glass, biotite, β — quartz). This suggests the beginning of an acid magma eruption.

The Upper Bourdigalian horizons are represented exclusively by terrigenous deposits: pebble beds, conglomerates, sands, sandstones, or motled green clays, red-purple sandy clays, often interbedded with thick beds and intercalations of rhyodacite tuffites. Their Bourdigalian age is inferred from their vertebrate fauna. (Vertebrates are known from northern Hungary, at Ipol Trnovts.)

Bourdigalian deposits are locally missing, and then Helvetian beds rest directly on various Chattian rocks.

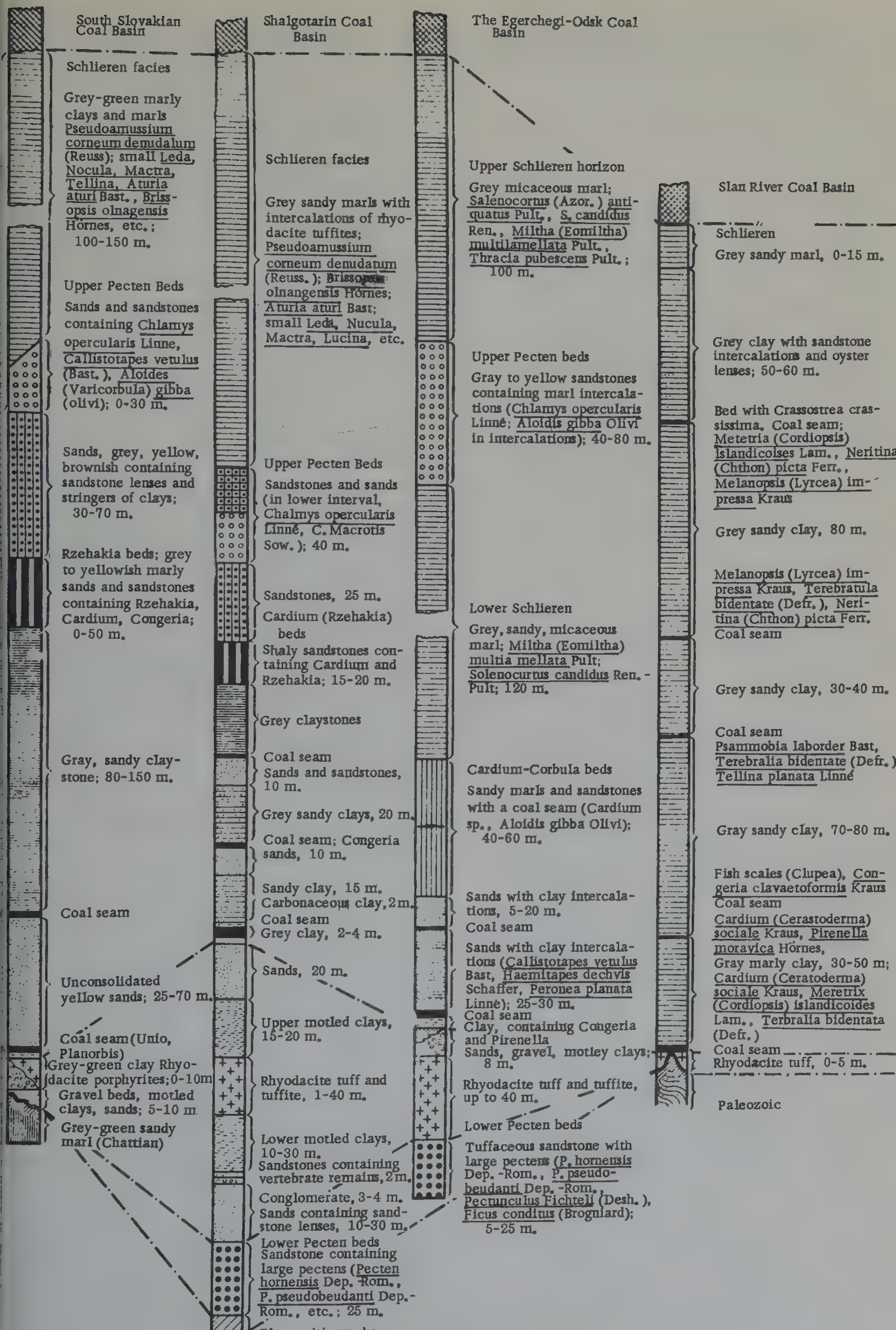
The Helvetian is represented here by a series of fresh-, brackish-water, and marine deposits, several hundred meters thick.

In the South Slovakian coal basin, the base of the Helvetian includes fresh- to slightly brackish-water clays and sands, and coal beds. Collected among the poorly preserved organic remains are Unio sp., Planorbis sp., and fragments of the bones of a small rhinoceros or tapir. Small congerias were found in drill cores obtained from the northern part of the basin. The producing layers are 25-70 m thick.

Present at the top of these beds are fresh- to brackish-water claystones, 80-150 m thick, barren of fossils except for a single exposure of obscure impressions reminiscent of small Cardidiae, and of redeposited foraminifera and sponge spicules. These clays gradually change to Rzehakia beds deposited between the fresh-water and marine sediments.

The distribution area of the Rzehakia beds in the South Slovakian coal basin can be established only approximately because its northern boundary is covered by younger Helvetian and Tortonian deposits while the south boundary is partially eroded. As determined by geological surveying and drilling, this area is about 250-300 square km. The thickness varies from 0 to 50 m. Lithologically and petrographically they are fairly uniform, being represented by light-grey, very fine-grained calcareous sands, highly micaceous, with sandstone lenses, and changing locally to siltstones. Irregular bedding has been observed. The carbonate content is about 14%.

Rzehakias (R. socialis Rzeh.) occur in large numbers, along with small Cardidiae, in layers 10-15 m thick; individual shells are scattered throughout the sand. The ratio of rzehakias to cardids changes from place to place, with the cardids alone present locally. A poorly



Stratigraphic position of the Rzehakia beds in the northern part of the Pannonian basin.

BM - Bourdigalian, marine facies; BC - Bourdigalian, continental facies; HFBM - Helvetian, fresh-water, brackish, and marine facies; TM - Tortonian, marine, often tuffaceous facies.

preserved Congeria ex gr. amygdaloides was collected in the vicinity of the village of Malyy Krtish.

Both the rzehakias and cardids often occur with their valves closed or open, which indicates their autochtonous origin.

The microfauna of the Rzehakia beds was recently described in detail from several drill-cores and exposures in the South Slovakian coal basin by K. Slaviková (1956), who noted its peculiar character. It consists of stunted foraminifera, no larger than 0. 05 mm, and of very small echinoid spicules. The most common foraminifera are Globigerina ex gr. bulloides d'Orb. and Rotalia beccarii (L.,), with less common representatives of the genera Cibicides, Bolivina, Bulimina, Elphidium, Nonion, and Cassidulina, and rare representatives of the genera Robulus, Lagena, Seiphonodosaria, Valvulineria, etc. The foraminifera are evenly distributed throughout the Rzehakias beds; the shell surfaces are so well preserved as to leave no doubt to their being autochtonous.

K. Slavnikova correctly infers from her ecological analysis that the Rzehakia beds were deposited in a brachy-haline marine, or a plagio-haline, brackish water basin, having a salinity of about 16%. She ascribes the origin of such a basin to the ingression of the Helvetian sea into the coastal lakes or lagoons, i. e., in agreement with our theory of the origin of Rzehakia beds in the Pannonian basin [25]. This assumption is corroborated by the presence of plankton steno-haline foraminifera (Globigerina) in the Rzehakia beds' biocenosis, thus suggesting the proximity of an open sea [38]. R. Lehotayová and E. Brestenská, who have studied the microfauna of the South Slovakian coal basin, do not rule out the possibility of the Rzehakia beds' microfauna having originated in that of the neighboring Pecten beds and become stunted in the new and unfavorable unfavorable environment (oral communication).

The top layer of the Rzehakia beds changes gradually to fine- to medium-grained sands with lenses of sandstones and intercalations of grey-green sandy clays. These sands are not as well sorted as the Rzehakia, and their carbonate content drops sharply (0-14%). Nonetheless, their mineral composition is not much different from that of the Rzehakia beds. They often contain manganese concretions and blooms. These beds are virtually barren of fossils with rzehakias and cardids observed in a single locality (near the village of Nova Ves), in hard sandstones at their base. These beds are 30-70 m thick, and are overlain by yellow to grey-green fine-grained calcareous sandstones with a steno-haline fauna consisting of fairly typical pectenids (Chlamys opercularis L., Chlamys scabrella Lam.).

Comparatively recently after a revision of the rauna, I. Chepregin-Meznerich [22, 23] established a Helvetian age for the so-called "Upper Pecten Beds" of Hungarian geologists. K. Slavikova (1956) identified there a fairly rich foraminifera fauna. Occurring along with benthonic steno-haline representatives of the genera Cibicides and Bulimina are those of genera Elphidium and Nonion, suggesting a slight local salinity decrease. Planktonic foraminifers are represented by Globigerina ex gr. bulloides d'Orb. or by Globigerina concinna Reuss. The Helvetian section culminates in fairly thick (100-150 m) grey to gray-green slaty sandy marls and marly clays of the schlieren facies. They contain a fairly rich, comparatively deep-water, fauna of the schlieren type: Pseudoamussium denudatum Reuss, Aturia aturi Bast., small tellins, nuculas, lucinids, mactras, ledas, etc. The microfauna, too, is quite rich in genera and species. The marls locally become sandy and the aspect of the fauna is somewhat different (appearance of simple corals). Following a break and denudation, the Helvetian was overlain mainly by tuffaceous deposits containing a rich Tortonian fauna. It should be noted that the character of the Helvetian deposits in the South Slovakian basin frequently changed; some of the beds are inconsistent in their thickness and locally are missing altogether (especially the Rzehakia and Pecten beds). For instance, in the vicinity of the village of Trench (Gushchin Hill), northeast of Modryy Kamen, the Rzehakia beds are replaced by Pecten sands, while the top of the fresh-water clays includes sands containing Chlamys opercularis L., small Callistotapes vetelus Bast., Aloidis gibba Olivi.), etc.

The Shalgotarjan coal basin, located approximately 40 km southeast of the South Slovakian basin in Hungary has been fairly well explored from cores drilled by the Uglerazvedka and presents an extension of the South Slovakian basin. Despite a few facies changes, the Shalgotarjan basin maintains, on the whole, the geologic structure of the South Slovakian Basin. Its facies changes are expressed in a thicker Bourdigalian terrigenous layer, thinner claystones in the top of the coal bed, and a considerably greater thickness of the Upper Helvetian schlieren facies (up to 300 m). It also appears that the producing beds were deposited in the basin of a somewhat different salinity: Teredoes appear in the third (Teredo) bed, following the mass appearance of Congeria ex gr. amygdaloides in the second Congeria bed. Here, too, the Rzehakia beds ("Cardium beds" of Hungarian geologists) appear between the brackish and marine deposits; as in the South Slovakian basin, they do not seem to have been developed over an area of any size. Their thickness, too, is only 15-20 m.

They are missing in some localities, such as

ır the village of Ipelsk Tyrnovets, at the echoslovakian border, where beds contain- ; typical Helvetian pectens are present at the of the brackish-water clays.

Still farther east, the Egerchegi-Ozdsk coal sin is located near the Bjukk Mountain foot- lls. As a result of denudation, the coal basin separated from the Shalgotarjan basin by a t of Upper Oligocene deposits. As in the preceding basins, the Helvetian deposits cur here between the Upper Bourdigalian faceous terrigenous rocks and transgressive rtonian deposits. Here, however, the change the Helvetian facies is considerably greater. st, the thickness of the producing layer and number of coal seams are reduced. The oductive formation retains brackish-water ına (Pirenella, Congeria) but more steno- ine genera, Callistotapes, Hemitapes, and llina appear in its upper beds. The transi- n to the true marine facies is marked by ıdy marls and sandstones (the so-called ardium-Corbula beds" of Hungarian geo- ists), Aloidis gibba and small cardids, nilar to those from the Rzehakia beds, which e 40-60 m thick here and contain occasional ritella sp. [37].

These beds are overlain by fairly thick rls and marly clays of the schlieren facies, ıtaining the arenoceous-argillaceous "Upper cten Beds", 40-60 m thick, and containing a ical fauna of Helvetian pectinids. (Chlamys ercularis L., etc.). The Pecten beds divide Schlieren facies into "lower Schlieren" and pper Schlieren", with similar faunas omiltha multilamellata Desh, Solecurtus iquatus Pult, etc.). The Helvetian is dis- ıformably overlain by Tortonian tuffaceous posits.

This description, together with the cross- ction, shows that while the Helvetian deposits ain their character, on the whole, the ehakia beds are likely to be missing in the ction. It is not impossible that their strati- aphic equivalent is represented by the ardium-Corbula beds" which, too, were posited in less saline waters.

While Helvetian deposits in the first three sins described have much in common, this not true for the fourth basin located in the ın River valley northeast of the Egerchegi- dsk basin. Here, the Bourdigalian and rtonian are separated by fairly thick and hologically uniform sandy clays containing al seams and with alternating, more or less esh water facies. The producing formation locally overlain by thin marls and marly clays hlieren). It is of interest that the lower part these deposits contain Cardium (Cerasto- rma) socialis Kraus, known also from the rchberg beds, as well as Pirenella moravica örn.), known from Moravia. In addition, Ostrea (Crassostrea) crassissima Lam. appears in mass, in oyster beds and lenses of the upper intervals. This lends support to R. L. Merklin's views [13] that the Rzehakia beds and those having the abundance of oysters are "facies of adjacent biotopes".

It is extremely important that, according to Z. Schréter [36], shallow brackish- to marine Helvetian facies are altogether missing on the northern slopes of the Matra Mountains south of the areas here described; out there, marls of the schlieren facies rest directly on the equivalents of fresh-water productive beds of the Shalgotarjan basin. It is, therefore, reasonable to assume that the calcareous member of the Helvetian lower schlieren facies is correlative with the brackish and shallow-water marine deposits along the margins of the basin.

On the basis of these data, the development of the Rzehakia beds in the northern part of the Pannonian basin is as follows.

Following the Upper Bourdigalian sea regression and land formation, a new subsidence of the Carpathian foothill trough resulted in another transgression, varying from place to place. Its initial stages were fairly rapid, becoming local ingressions. As a result, comparatively deep-water schlieren facies were deposited directly on the fresh-water (lacustrine) beds correlative with the Shalgotarjan producing series. As the coastal plain continued to subside, water filled the coastal depressions, forming first lakes then lagoons, gradually changing to straits and embayments. Later, the lagune and lacustrine deposits were buried first under shallow- then under deeper marine deposits (schlieren). The Rzehakia beds were deposited in lagoons connected with the open sea, as evidenced by the presence in them of planktonic steno-haline foraminifera. Rzehakias occur in the shallow-water calcareous micaceous sands and silts, more or less sandy or silty argillaceous, deposited in comparatively shallow waters of lower salinity and agitated by weak currents. Any marked change in the granulometric composition of the sand or in the salinity of water caused the Rzehakia to disappear. This is the explanation of the fact that the Rzehakia beds of the Pannonian basin are present locally only and are never very thick.

CONCLUSIONS

1. Rzehakias are inhabitants of brackish waters, and their association with a marine steno-haline fauna suggests their synchronous or asynchronous redeposition. It is always possible that these steno-halite genera present in a Rzehakia biocenosis are relict forms, often stunted; as such, they are of a secondary

importance. It also should be kept in mind that Rzehakia beds of many provinces were eroded by subsequent sea transgressions.

As a rule, rzehakias appear in transitions from marine to fresh-water conditions, or vice versa.

2. Rzehakia beds were deposited in foredeeps as well as in interior troughs. In the first, they were concentrated largely along the axis and around the outer edges, toward the platform. The inner margin of the foredeep, near the rising and advancing mountain ranges, greater depths and steeper shores developed, with a resulting great influx of coarse clastic material. Such conditions were not conducive to the development of any fauna, let alone a Rzehakia fauna which is very sensitive to ecological changes.

3. The deposition sites of Rzehakia beds were marine areas of reduced salinity, such as embayments gradually cut off from the open sea and replenished by rivers. The deltas and estuaries of these rivers separated the embayments into relict lagoons. Rzehakia beds could also be deposited in depression lagoons formed by the sea ingressing over a slightly hilly coastal plain in advance of a general inundation.

The transgressive position of Rzehakia beds directly above the older cosk in some areas of the foredeeps (for instance in South Germany and Switzerland) is explained by the fact that orogenic movements continued after local desalting of the foredeep, so that the lower salinity areas were shifted toward the platform.

4. The prevailing view was, until now, that Rzehakia beds had not been deposited prior to the Middle Miocene. The discovery of horizons containing small rzehakias in the Oligocene deposits of the southern U. S. S. R., kindly reported to us by R. L. Merklin, indicates their wider vertical range.

Admittedly, the data cited here are incomplete, especially on the phylogeny and biocenosis of Rzehakias. Their study from Oligocene deposits will undoubtedly clarify this problem.

The above exposition gives some idea of the difficulties in determining the origin of Rzehakia beds. A correct solution of this problem will advance the knowledge of Tertiary paleogeography and stratigraphy.

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BRIEF COMMUNICATIONS

NEW DATA ON THE TECTONICS OF THE APSHERON SHELF[1]

by

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This article describes the results of the 1957-1958 seismic work in the Caspian Sea. Materials resulting from this marine seismic study, along with data on the bottom relief, represent the first factual information from which to compile a true picture of the tectonics of the western part of the Apsheron shelf — a submarine bridge between the Apsheron oil and gas district on the west and the Balkhan area in the east.

The Apsheron shelf extends from the Apsheron peninsula westward to the Krasnovodsk-Cheleken shoreline of the east coast. It divides the submarine relief of the present day Caspian Sea into two uneven parts: the (larger) northern and the smaller but deeper (southern).

The association with the extremities of this shelf, of such immensely rich oil and gas provinces as the Apsheron in the west and the Balkhan in the east makes it possible to speak of oil and gas in the Apsheron shelf itself. This possibility is convincingly illustrated by the numerous mud volcanoes (Livanov, Gubkin, Zhdanov, etc., banks) and by sporadic eruptions in the Apsheron shelf area itself.

The geology of the Apsheron shelf is treated in the voluminous literature [1, 6, 10, 12, 14], especially in respect to the relationship between the Caucasian and Transcaspian orogenies [2, 3, 5, 9, 13, 15].

It should be noted, however, that until recently, there were practically no definite data on the tectonics of the Apsheron shelf proper, although it is these data that could have clarified the extremely controversial problem of the geologic relationships between the Caucasus and the Transcaspian.

The pendulum and aeromagnetic survey data, together with those on bottom relief — valuable as they are — do not afford means for a quantitative and geometrical presentation of the results [7, 12, 17, 18]. Such means is provided by seismic exploration.

Prior to 1957, seismic work was carried out only in the immediately offshore parts of the shelf, with depths not exceeding 70-100 m. In 1957, a study of the Apsheron shelf tectonics was initiated simultaneously on both sides: west of the Apsheron Archipelago, by the Azneft marine seismic party; and from the Turkmenian side in the east — by the All-Union Scientific-Research Geological Institute party. Using the recently-developed floating system of receiving stations ("pieze-bar"), the eastern party ran, in addition to profiles in the Cheleken area, a seismic profile from the Livanov Banks to Neftyanyye Kamni. In 1957-1958, an Azerbaydzhan seismic party covered an area of about 200 square kilometers with a reconnaissance network, supplemented locally by a semi-detailed profile network.

As a result, new anticlinal structures have been discovered east of Neftyanyye Kamni: The April 28 Anticline, and the 26 Baku Commissars' Anticline. An idea of these uplifts can be obtained from the structural sketch drawn on a conditional seismic horizon associated with the Sabunchinsk producing formation (Figure 1), also from typical seismic profiles crossing them and the adjacent areas of the Apsheron shelf in two directions normal to each other (Figure 2). These anticlinal structures have a northwest-southeast (Caucasian) trend and are located in a tectonic zone whose western part contains such deeply drilled structures as the Zhiloy Island fold and Neftyanyye Kamni. The newly-discovered structures are fairly strongly disrupted and asymmetric; the seismic survey reveals the presence of faults near and at the crests.

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[1] Novyye dannyye o tektonike Apsheronskogo Poroga, (pp. 100-105).

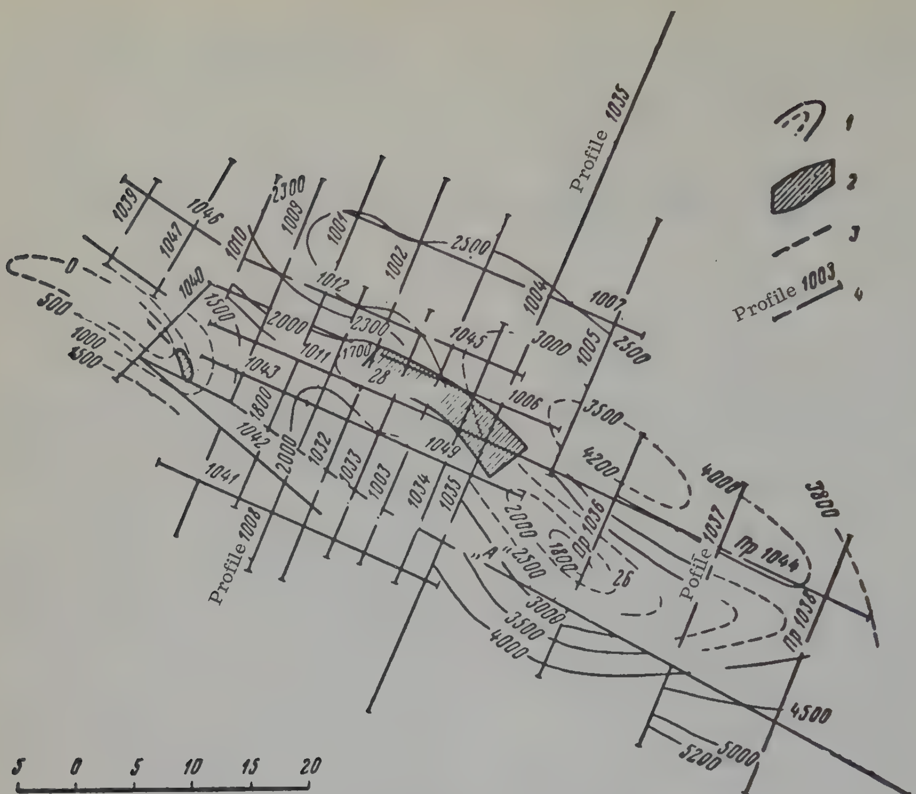

FIGURE 1. A structural map drawn on a conditional horizon (Sabuchinsk formation of the producing stratum) in the northwestern part of the Apsheron shelf

1 - isolines of the conditional seismic horizon; 2 - inferred fault zone from seismic data; 3 - fault discovered by drilling; 4 - seismic profiles.
The structure of the 26 Baku Commissars' anticline is not tied in with the regional structure (a thinly-spaced profile network and the possible presence of faults between the April 28 and the 26 Baku Commissars' anticlines).

The nature of the junction of the two structures is not clear: deformations are possible also in the saddle between them. These structures are fringed by deep and fairly extensive synclines; their proximity to the already known structures in the southeastern part of the Apsheron Archipelago (Neftyanyye Kamni, etc.) and the community of their geologic structure make it possible to assign this western part of the Apsheron shelf to a geosynclinal province.

The proximity of the newly-discovered uplifts to the Apsheron oil and gas province to its productive formation suggests the possibility for their productive potential. The depth of the top of the Sabunchinsk productive formation in the crests of these uplifts indicates a considerable subsidence of this part of the tectonic zone, as compared with its northwestern part — a component of the Apsheron Archipelago folding. This excessive depth complicates exploration and drilling; on the other hand, the considerable thickness of younger deposits (Akchagylian, Apsheronian, and Quaternary) over this potentially productive formation may act as a cap rock for its oil and gas.

It should be noted, however, that these structures lie under 100-200 m of water.

Some seismic profiles run by the Azerbaydzhan seismic party in 1957-1958 suggest the presence of still other structures. Almost all southwest-northeast profiles, northeast of the April 28 and the 26 Baku Commissars' anticlinal flexure show a deep asymmetric syncline whose northeastern limb forms a gentle homocline. The longest profile (Figure 2) shows still another large asymmetric anticlinal flexure beyond that; it is comparatively flat along this profile.

The presence of such a flexure suggest two hypotheses of equal value as yet from which to choose. It may be that this flexure is a periclinal segment of a geosynclinal fold (or a

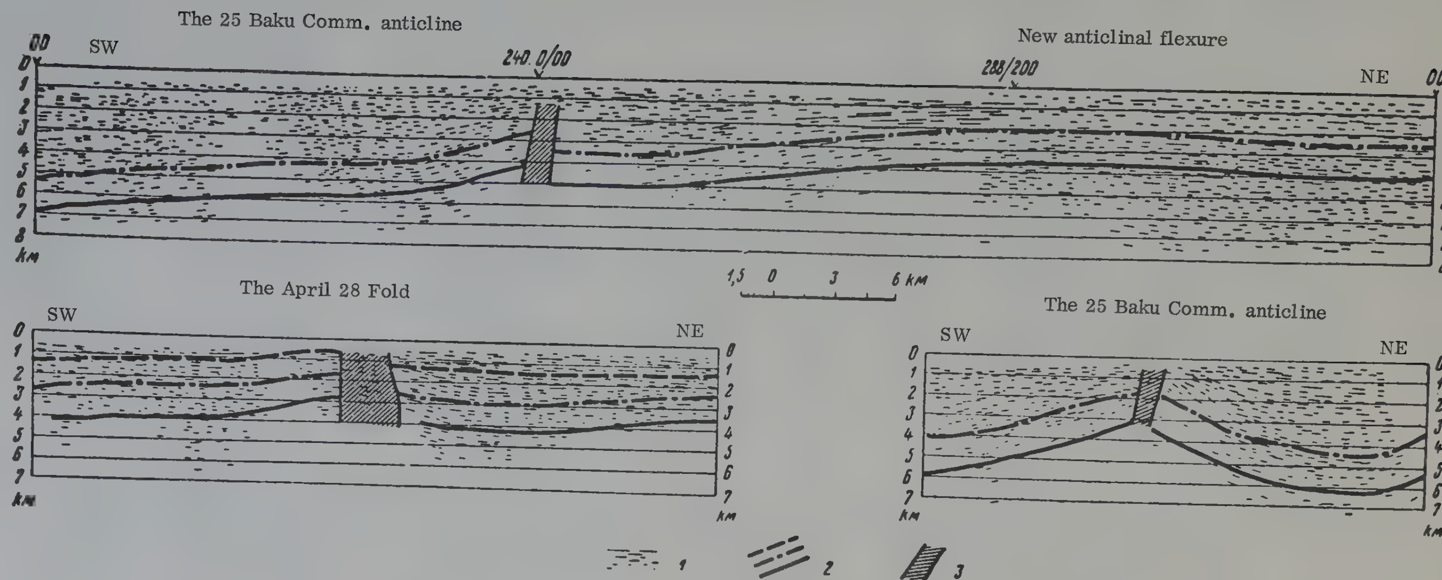

FIGURE 2. Seismic profiles in the southwestern part of the Apsheron shelf

1 - reflecting areas; 2 - various conditional seismic horizons; 3 - faults zones inferred from seismic data.

system of folds) similar to those in the Apsheron Archipelago (such as the Neftyanyye Kamni or the 26 Baku Commissars' folds). In this case, a steep fold may lie northwest or southeast of that flexure (assuming the Caucasian trend of folding), or else on both sides of it. Naturally, the presence of such a fold (or a system of folds) substantially enlarges the geosynclinal part of the Apsheron shelf, pushing its northern boundary toward greater depths. This would increase the oil and gas potential of the northern part of the shelf, insofar as it is probable that the producing formation here may occur at shallower depths than in the southern part.

On the other hand, it is possible that the anticlinal structure of folds in the northern part is altered because of its relative proximity to the platform province, the so-called hypothetical buried North Caspian land (Figure 3). As is well known, the existence of a platform in the central part of the Caspian Sea is suggested by the data of pendulum and aeromagnetic surveyings [7, 18]; however, all published tecttonic maps show its southern edge to be somewhat northeast of the area of the 1957-1958 seismic work [17, 18, 20].

importance in this respect are the seismic operations being carried on in the Krasnodar-Cheleken coastal area, by the All-Union Scientific-Research Geophysical Institute. Even less is known on the tectonics north of the Apsheron shelf.

It is to be hoped that future seismic work in the central and south parts of the Caspian Sea will result in comprehensive material on the tectonics of the Apsheron shelf proper as well as on its relation to the adjacent tectonic features.

It will shed light on the V. F. Solov'yev assertion that the Apsheron shelf is not a single geologic unit but "is rather to be regarded as the junction of three structural features: the subsidence area of the southeastern Caucasus Tertiary folding; the Krasnovodsk peninsula structural area (Paleozoic platform with a younger sedimentary mantle); and the Tertiary folding province of the West Turkmenian trough" [17].

The results of the 1957-1958 work in the Caspian Sea suggest the necessity of further

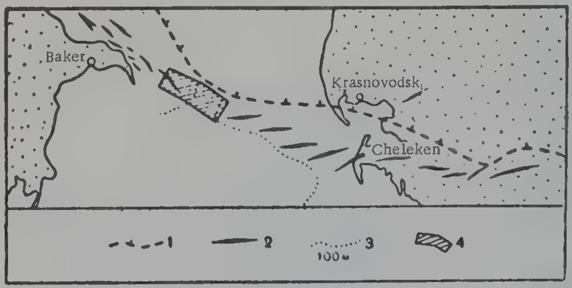

FIGURE 3. Tectonic sketch of the Apsheron shelf (slightly modified after V.Ye. Khain, and others [20]).

1 - boundary between the Paleozoic platform and the younger sedimentary mantle; 2 - axes of anticlinal structures; 3 - isobaths; 4 - area of the 1957-1959 seismic work.

A seismic determination of the southern boundary of this platform province and of the tectonic nature of its junction with the geosynclinal province may have an important bearing on the approach to the problem of the source of sediments now including the producing formation — a subject of a lively debate in geological literature [4, 8, 11, 19, 20].

Thus, the deep-water seismic work of 1957-1958 has established a rather close tectonic connection between the Apsheron Archipelago and the western segment of the Apsheron shelf.

The nature of the folding farther east and northeast of the 1957-1958 study area is difficult to visualize at present. Of considerable

seismic study in its central and southern parts to determine their tectonics and the possible presence of new oil- and gas-bearing structures.

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SPHERULITE LAVAS FROM THE VICINITY OF GAMZACHIMAN VILLAGE[2]

by

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Spherulitic inclusions may supply a key to understanding the processes taking place in a magma prior to its crystallization. This explains the interest shown in them by a number of investigators.

These formations were described in basic rocks as early as the end of the last century, by Ye. Delles, A. A. Inostrantsev [5], I. I. Tolmachev [10], and subsequently by F. Yu. Levinson-Lessing [7-9].

In recent years, in connection with the discovery of spherical bodies in acid rocks, interest in them has been shown by T. L. Tanton [11], D. S. Belyankin [1, 2], I. M. Volovikova [3], V. I. Lebedyanskiy [6], and O. P. Yeliseyeva [4]. Most of those authors ascribed the origin of these bodies to the liquation process.

This article deals with the spherical inclusions in the acid lavas from the south slopes of the Bazum Range (Kirovakanskiy rayon, Armenian Republic).

Geologically, this area is part of a broad expanse of Paleogene deposits — the Sevan-Shirak synclinorium or Sevan tectonic zone.

The south slopes of the Bazum Range are composed of Middle- to Upper Eocene volcanic and volcanic-sedimentary rocks on the south limb of the Khalab anticline.

The Middle Eocene deposits (total thickness, about 2000 m) are represented by hornblende-, augite-, and plagioclase andesites, their tuffs, and tuffobreccias with subordinate tuffaceous sandstones and conglomerates. Some investigators assign to the top of the Middle Eocene a strongly eroded formation of keratophyres, their tuffs, and tuffobreccias.

Resting on various Middle Eocene horizons, on a sharp unconformity and a basal conglomerate, are the Upper Eocene volcanic formations (up to 1000 m thick) which comprise the Bazum Range divide. The bottom of the Upper Eocene begins with tuffaceous sandstones and tuffobreccias interbedded with andesites and andesite-basalts; above these are acid to alkalic effusives, their tuffs, and tuffobreccias. The Upper Eocene section culminates in liparite-dacite extrusions. Still higher (the Dilizhan River area) are Oligocene to Lower Miocene fresh-water lacustrine deposits.

The spheroidal bodies were observed in one of these Upper Eocene extrusives exposed 5 km north of the village of Gamzachiman near the Kzlarkhach nomad settlement. This extrusion is irregular, measuring 2x0. 5 km, and elogated to the northwest. It is exposed in the right side of a gorge west of the Tar-Dara summit (Figure 1).

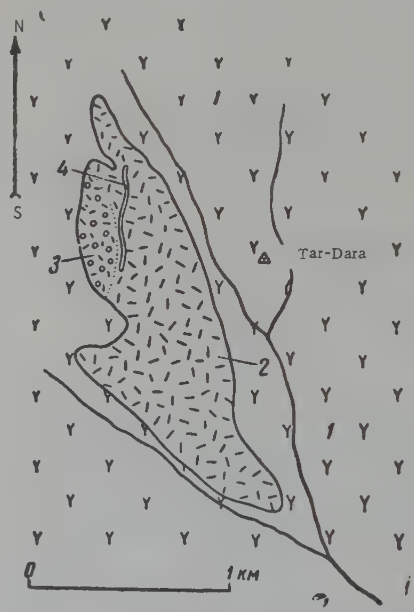

FIGURE 1. Sketch map of the Gamzachiman village extrusion

1 - the enclosing andesites; 2 - liparites and liparite-dacites; 3 - spherulitic lavas; 4 - felsite dike

The extrusive body consists of pink-grey to pink-purple, liparitic to liparite-dacite rocks showing flow structure, and contain large (up to 0. 5 cm) and rare feldspar incrustations. No definite orientation in the flow structure has been observed except for a general steep dip of the flow lamellae. As a result of weathering, parts of the extrusion stand out in sharp and bizarre peaks up to 35-50 m high. The north part of the extrusion is cut by a felsite dike, trending meridionally and dipping west at 75-80°. This dike is up to 15 m thick and is traceable for 400 m, conspicuous in its columnar jointing and is a light-grey color.

Spherulites were observed in the western,

[2]Sferolitovyye lavy okrestnostey sela Gamzachiman, (pp. 105-110).

and topographically highest, part of the body, in an uneven area. Sections containing these spherical bodies stand out from the other rocks: the homogeneous pink-grey liparites change here to strongly differentiated formations consisting of a blue-grey ground mass and pink-grey spherules (Figure 2).

spherules are less distinctive, being marked by greenish to brown-green spots with haphazardly-arranged transparent nuclei; they are more conspicuous when seen with crossed Nicols. The spots are vitreous, poorly-polarizing areas, while the transparent nuclei are recystallization centers.

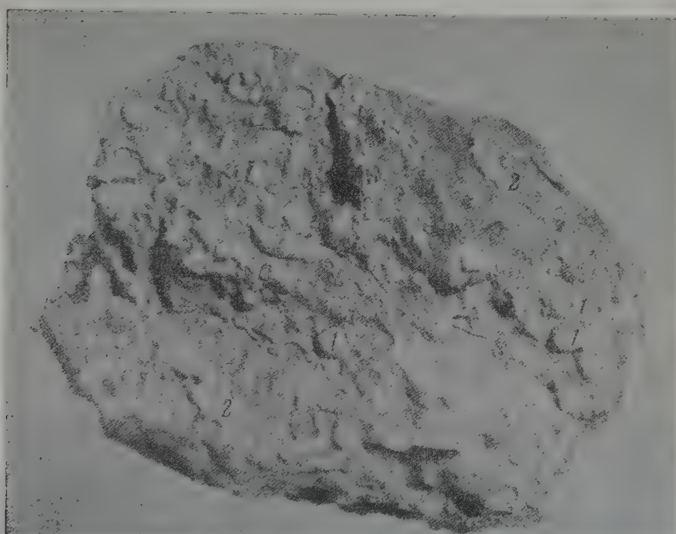

FIGURE 2. Spherulite lava

1 - spherulites; 2 - cementing groundmass.

The spherules are unevenly distributed, giving the impression that they are associated with definite planes where they account for 40-50% of the rock and almost wholly displace the groundmass. They are in sharp contact with the latter and are readily removed with the hammer. They often form reniform of "warty" aggregates. Their size varies from 0. 1 to 2. 5 cm, the most common diameters being 0. 5-1. 0 cm.

In thin sections, the liperaites show porphyritic structure with elongated incrustations of oligoclase, 1. 8-2. 0 mm. The quartz-feldspathic groundmass is very slightly felsitic and poorly polarizing.

The spherule-containing rocks are represented by a very finely crystalline groundmass in which the spherules are immersed.

The cementing groundmass is quartz-feldspathic with rare microlites of plagioclase, "streamlining" the spherules. The latter are of several varieties associated with transition.

The vitreous spherules differ sharply from the cementing groundmass. The crystallized

Some spherules exhibit a radial structure, becoming "washed out" away from the center, often changing to a concentric, almost isotropic fringe up to 0. 02 mm thick. Other spherules are felsitic, differing from the cementing groundmass by finer crystallization. The larger spherules contain, at times, smaller sphero-crystals (Figure 3). All spheres consist of glass, its crystallization products, crystalline inclusions of feldspar, and ore mineral incrustations.

Their index of refraction is higher than that of Canada balsam.

In addition to these spherules, the thin sections occasionally show typical sphero-crystals. They differ from the spherules in: 1) a conspicuous growth center; 2B) a well-developed radial structure (radial crystals standing out sharply against the cementing groundmass); and 3) in serrate outlines, apparently formed by uneven growth of the crystals (Figure 4).

The distribution of feldspathic incrustations shows no relation to the spherules, occurring in, out, and cutting them.

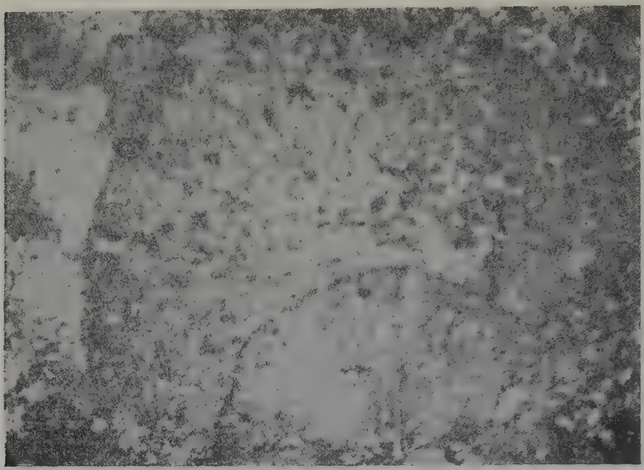

FIGURE 3. A segment of spherocrystallization with spherulites

Magnification X32; crossed Nicols.

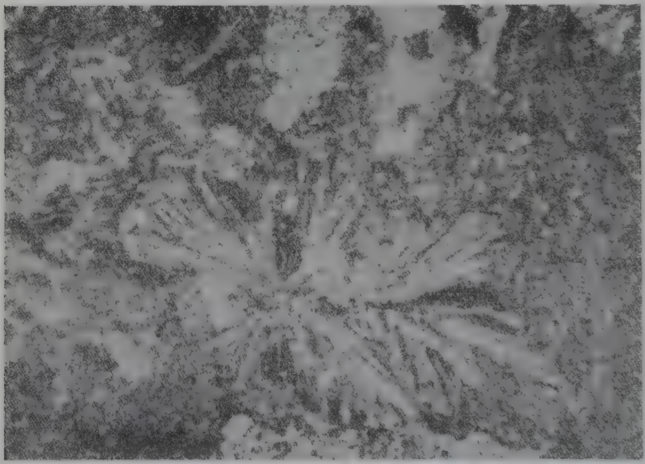

FIGURE 4. Serrate outlines of spherocrystals

Magnification X32; crossed Nicols.

Table 1 lists the chemical analysis of the gross sample of the spherules and cementing groundmass.[3]

The analysis shows that the spherules differ from the groundmass in lower SiO_2 content, a higher alumina content, and an appreciable increase in alkalis.

In A. N. Zavaritskiy's diagram, the representative poing for spherulites is close to that for quartz latite, the difference being in its higher SiO_2 content and the lower total ferric and ferrous iron. The representative point for the cementing groundmass, on the other hand, corresponds to the intermediate liparite of R. Daly, differing from it in a higher content of alumina and alkalis. The chemical analyses were converted to standard by the CYPW method.

There are several views concerning the origin of various spherules.

[3]Analysis made by M. M. Yazydzhyan at the Chemical Laboratory of the Institute of Geological Sciences, the Armenian S. S. R. Academy of Sciences.

Table 1

Oxides	Gross composition	Spherules	Groundmass	Characteristic	Anal. 513	Anal. 514	Anal. 515
	Anal. 513	Anal. 514	Anal. 515				
SiO_2	73.60	69.79	78.95	a	14 4	13.8	9.7
TiO_2	0.30	0.21	0.27	c	0.8	0.8	0.4
Al_2O_3	12.70	17.33	10.89	b	4.2	9.8	5.3
Fe_2O_3	1.10	0.60	1.04	s	80.6	75.6	84.6
FeO	1.86	1.77	0.90	a'	9.4	67.6	63.4
MnO	0.04	0.07	0.03	f'	62.5	22.5	29.3
MgO	0.75	0.58	0.25	m'	28.1	9.9	7.3
CaO	0.66	0.64	0.35	c'	—	—	—
Na_2O	3.68	3.73	2.07	n	18.3	56.6	45.3
K_2O	4.75	4.30	3.91	φ	21.9	5.3	14.6
H_2O	0.25	0.46	0.24	t	0.3	0.3	0.3
Loss in heating	0.60	0 64	0.69	Q	31.6	22.8	49.4
F	Not det'd.	0.55	0.58	a/c	18.0	17.2	24.0
Cl	—	Trace	Trace	—	—	—	—
S	—	0.06	»	—	—	—	—
P_2O_5	—	Trace	—	—	—	—	—

Table 2

Minerals	Spherules, sample 514, in %	Groundmass, sample 515, in %
Quartz	31.59	51.41
Orthoclase	25.60	22.82
Albite	31.46	17.82
Hypersthene	3.41	1.05
Corundum	6.52	3.26
Magnetite	0.93	0.61
Fluorite	0.94	0.47
Ilmenite	0.46	0.79

A number of authors, in stressing the well-developed radial structure of certain spherules, believe them to be the result of spherulitic crystallization from a melt. As early as 1874, A. A. Inostrantsev regarded as a criterion of such crystallization, the appearance of a cross pattern in polarized light taking in the entire spherule.

Other authors regard these spherules as droplets of a liquated magma and point to their felsitic vitreous structure and their form — not always spherical and often irregular. Still others associate the formation of these spherules with a later crystallization of a vitreous mass; finally there are those who believe that all of these ways of spherulite formation are possible.

Data cited in this article suggest that the spherules in the acid lavas of the Kirovakan area were formed, with a few exceptions, by liquation. The supporting evidence is as follows:

1. The spherules occur in various stages of crystallization, as witness their variable and transitional structures — from vitreous to very finely felsitic to radial. Even in radial structures, the spherules show a concentric vitreous fringe.

2. The spherules are mostly spherical to oblate, with smooth outlines as against the serrate spherocrystals.

3. Individual spherules often occur in reniform and warty growths or in zones; the latter suggest their participation in some movement. This, in turn, suggests their presence in a melt, simultaneously with the cementing groundmass.

4. The spherules differ sharply from the groundmass, in their content of principal oxides and in a higher amount of such elements as manganese, chromium, zirconium, silver, and gallium (as determined by the Spectroscopic Laboratory of the Institute of Geologic Sciences, Academy of Sciences, Armenian SSR; G. M. Mkrtchyan and M. Ya. Martirosyan, analysts).

5. The diversified and complex standard mineral composition of the spherules and the groundmass militates against a sphero-crystallization or a partial crystallization of glass, because the latter contains usually only one or two minerals.

The study of these spherules suggests their

formation mechanism as indicated by F. Yu. Levinson-Lessing, viz. physico-chemical conditions favorable for differentiating a homogeneous lava into two immiscible melts were created by the assimilation process in the presence of volatiles (water, fluorine, chlorine, and sulfur, in our instance).

The streamlined arrangement of these spherules suggests that this differentiation took place at some depth, with subsequent displacements. Our data are inadequate to estimate the magnitude of these displacements.

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REVIEWS AND DISCUSSIONS

A FEW OBSERVATIONS ON I. M. SUKHOV'S ARTICLE, "ON THE AGE OF NON-FOSSILIFEROUS LOWER PALEOZOIC SECTIONS IN THE DNESTR REGION."[1,2]

by

D. Ye. Panchenko

The classic Lower Paleozoic section in the Dnepr region has engaged the attention of geologists for a long time. While everybody agrees on the Ordovician and Silurian age of its upper terrigenous carbonate part of the section which has an abundant fauna, the age of the lower part of the section which is virtually non-fossiliferous is still a subject for argument. In estimating the age of this section of the section, there is still wide disagreement on the age of any layer in the section, certain geologists relating it to the Ordovician; others to the Cambrian, and still others to the Riphean, and Precambrian.

I. M. Sukhov holds the extreme view of T. Văscăutnu, as early as 1931, according to which the entire section, beginning with the arkose member, is Ordovician.

It is known that the Văscăutnu's classification is based not so much on paleontologic data as on a pre-Silurian correlation of his own. He believed that what the Dnestr cuts is not a syncline, but a very large synclinorium with gently dipping limbs which he called the "Atak basin". According to him, the central part of this basin is filled up with argillaceous material (Naslavchi shales); the eastern coastal area, with the Atak sandstones; and the western, with Molodovo sandstones. In his interpretation, the age of the Atak basin rocks was determined not from the scanty fossil remains observed by that author in the Naslavchi shales but rather from the Molodovo horizon, well represented paleontologically. On the subject of the Naslavchi fossils themselves, T. Văscăutnu states that "not a single stratigraphically important form has been found among them".[3]

Although the concept of a single Atak basin has lost all of its advocates, there are those who still hold to an Ordovician age for ancient formations of the Dnestr region. Among them, in addition to Sukhov, are V. P. Kurochka. A. Ya. Edel'shteyn, and others. They refer to the above-mentioned classification and data by T. Văscăutnu, overlooking the fact that Văscăutnu himself had no evidence other than the Molodovo horizon fauna. Now, it is well known that this horizon is not a facies of the phosphoritic shale section; moreover, a long interval separates it from the underlying terrigenous sequence.

By now the reader is well aware of Sukhov's error in interpreting the Văscăutnu's stratigraphic section. In his brief discussion on p. 395, the author completely distorts its meaning and the Văscăutnu correlation of Atak sandstones, Naslavchi shales, and Molodovo sandstones. In an appended table, Sukhov attempts, again unsuccessfully, to set the reader straight in a correct interpretation of that section.

What the table does show is Sukhov's lack of familiarity with the literature on the correlation of ancient sequences.

We have no way of judging the degree of preservation of the microfauna found by Văscăutnu in the Naslavchi shales, much less his identification. However, it is reasonable to suppose that he could not have identified a trilobite cranidium fragment as anything else than an Asaphus fragment, inasmuch as he was convinced that he was dealing with Ordovician fossils. Nonetheless, he attached no stratigraphic importance to these findings, while

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[1] Nekotoriye zamechaniya po povodu stat'i I. M. Sukhova "O vozraste nemykh tolshch Nizhnego Paleozova v Pridnestrov'ye", (pp. 111-113).

[2] Doklady Akad. Nauk SSSR, t. 124, no. 2, 1959

[3] T. Vascautnu's works are cited from Spellman's (?) translation.

Sukhov et al. magnify its importance and make it the basis of their speculations.

Sukhov states that the basis of his stratigraphic scheme is the orderly sequence of arenaceous and argillaceous horizons, i. e., cyclic sedimentation. In his scheme, a cycle constitutes a subformation, which is differentiated into horizons. He borrows the proprietary names for his stratigraphic units from earlier authors, mainly from M. F. Stashchuk [5].

We cannot help but note that the names of his formations and subformations are unfortunate. For instance, he applies the old name, "Ushitsa formation", to a new deposit.

In grouping only horizons and subformations according to Stashchuk's section, i. e., cyclic sedimentation. But Stashchuk is known to have been guided not by the granulometric composition of the rocks which here are quite variable, but by their color. As a result, Sukhov's subformation of the Ushitsa formations turned out to be in disagreement with the principle of this classification — the cyclicity of sedimentation.

Withal, the age of these rocks is still the least substantiated point in Sukhov's stratigraphic scheme, inasmuch as even his most recent data can not be used as a reliable criterion on which to assign an Ordovician age to the entire section, let alone its differentiation into stages.

Citing, along with Văscăutnu, the evidence of the Molodovo horizon, Sukhov correctly rejects the idea of an "Atak basin". However, he warns the reader that "it should not be overlooked that, judging from the nature of the sediments, the breaks between the Molodovo (Upper Ordovician) sandstones and underlying sections, as well as between the other formations, were comparatively short" (p. 396). The author does not even attempt to substantiate this purely speculative conclusion. Therefore, both he and the reader will have a difficult time in explaining how such a short pre-Molodovo break could bring about such substantial changes in the paleogeography of the basin. On the other hand, what guidance did Sukhov use in differentiating his Ordovician into stages? He has abandoned the Văscăuťnu scheme, and there are no adequate data for a new one. So he "deduced" these stages from a history of the Ordovician basin as conceived by himself, with no relation to the known basin development on the Russian platform [1].

Having excluded the Valday deposits from the Dnester section, Sukhov has taken them out, for good measure, of the entire Bessarabian section, by assuming the presence here of continental conditions. However, this conclusion is contrary to the field data of A. G. Zavidonova [2] and other authors, indicating the presence not only of the Valday but also of Cambrian deposits in the ancient section of Bessarabia.

Finally there are the still unpublished paleontologic data of N. N. Sludskiy on Atak sandstones. Indeed, the fossils are scarce in these ancient formations; therefore, every new fact must be most carefully scrutinized. Unfortunately, the only new discovery is a poorly preserved impression reminiscent of Dictyonema sp., which does not resolve the argument about the age of these Dnestr region rocks. These data are just as valid for either the Ordovician or Cambrian. Considering the opinion of L. N. Repina and V. V. Khomentovskiy [4] to the effect that our knowledge of fossils in ancient beds is extremely limited and that there are findings of "highly organized forms... in deposits definitely older than Lower Cambrian", there is nothing surprising in discovery of the fossil remains of organic animals in these ancient rocks.

As of now, a great majority of students agree that there are some Lower Cambrian and Riphean deposits among the non-fossiliferous rocks of the Denstr region. The author of the work reviewed is alone in his consistent adherence to the old views about these formations. Determining their absolute age may lead to a rapprochement, if not to unity of opinion.

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ON N. D. SOBOLEV'S ARTICLE, "NEIVITE, A NEW VEIN ROCK"[4]

by

A. A. Spasskiy, and N. A. Temikov

N. D. Sobolev's article in No. 10, 1959, of this journal, describes an alleged variety of vein rocks in the East Tagil ultrabasic massif [4], which the author regards as a new rock and gives it the new name, "neivite".

According to N. D. Sobolev, neivites are magmatic vein rocks consisting of varying amounts of primary albite and common brown hornblende, containing small amounts of such secondary minerals as zoicite, epidote, and chlorite developed largely on the hornblende; there is up to 1% of accessory magnetite; very rare grains of sphene and apatite are also present.

The author's arguments for a primary magmatic nature for neivite are as follows: 1) the presence of unaltered rocks free of secondary minerals; 2) conspicuous zonation of the albite; complete absence of any relict minerals and structures (p. 119). In conclusion, he states, "the decisive argument for the magmatic nature of these rocks is the presence of a second phase in veins cutting the first intrusive stage veins. The first phase rocks are altered at the contacts as well as in xenoliths... Had there been a subsequent albitization, uralitization, or any 'greenstone regeneration', these two phases would have been indistinguishable..." (p. 119).

As long as these vein rocks are genetically related to deposits of alkalic amphiboles of great interest mineralogically, we deem it pertinent to consider these data critically.

Vein rocks of this area consist of variable amounts of albite, hornblende, and of pyroxene (omitted by N. D. Sobolev); the amount of this last mineral is 0-25%. Secondary zoicite and epidote take up to 30-40% of the thin section area. The accessory minerals, as noted by N. D. Sobolev, are magnetite and grains of sphene and epidote. Plagioclase is present in amounts of up to 40-50%, largely in elongated prismatic crystals, up to 0.5 mm long, in standard structural types. Its zonation is occasionally visible under the microscope. In that event, the central part of the crystals is completely replaced, in most instances, by a dense aggregate of zoicite and epidote, so that it is impossible to determine the composition of the plagioclase. A determination by the Becke method and on Fedorov's stage, of those plagioclase grains which are more or less free of secondary minerals, has shown that they belong to albite-oligoclase, Nos. 3-14. The considerable content of secondary minerals, along with their replacement of the oligoclase, definitely demonstrates the secondary nature of this abite-oligoclase. The secondary nature of albite in vein rocks of this area was indicated as early as this by A. Ye. Malakhov [3], V. V. Arshinov [2] and Yu. K. Andreyev [1], who believed that the albitization had affected the plagioclase of an andesine and even labradorite type.

We are not convinced by N. D. Sobolev's arguments [4, p. 119] that the zoning of albite is evidence of its primary magmatic origin because, when andesine is partially replaced by albite, its original zonation remains intact.

Hornblende, present in amounts from 35 to 65%, is very characteristic of this rock. The hornblende is brown with greenish, and often reddish, pleochroic tints. Its color is similar to that in Proterozoic basic rocks, quite different from the hornblendes in diorite. Sobolev notes that this characteristic color is a beige-greenish blend [4]. The optical constants of this hornblende, as measured in a great number of sections, are identical with those determined by N. D. Sobolev, and characterize it as a common hornblende. Monoclinic pyroxene is quite common, occurring in amounts stated above, in growths containing hornblende or individual crystals, always showing their typical cleavage.

We have observed monoclinic pyroxene in

[4] Po povodu stat'i N. D. Soboleva "Neyvit-novaya gornaya poroda iz gruppy zhil'nykh, (pp. 113-114).

almost all thin sections — a considerable number of which were taken from localities studied by Sobolev.

According to its optical constants, it is augite, with $\gamma - \alpha = 0.023$-0.025; $c\gamma = 40$-$60°$; and 2V, from about +50° to +58°.

V. V. Arshinov, B. Ya. Merenkov, and Yu. K. Andreyev, also noted the presence of pyroxene in these rocks.

The relationships between the brown hornblende and the monoclinic pyroxene suggest an earlier origin of the latter, because the hornblende often forms a fringe around it or replaces it so that the pyroxene occurs in a hornblende grain. Undoubtedly, the hornblende has developed by replacing the pyroxene, probably during the magmatic stage of rock formation.

Secondary alterations of dark-colored minerals, whose presence is categorically denied by Sobolev, have been found to be present everywhere as replacements of hornblende and pyroxene by a fibrous blue-green hornblende.

As to the second intrusive phase of N. D. Sobolev's "decisive argument", it should be noted that the second phase rocks differ sharply from the first, in structure. This is a very fine-grained rock, reminiscent of odinite, and consisting of a network of fine grains of hornblende and plagioclase, with porphyritic incrustations of monoclinic pyroxene scattered among them. It is reasonable to suppose that because of this fine grain, the second phase rocks will stand out against the coarse-grained first phase rocks in the subsequent greenstone alteration — just as is the case.

This brief description shows that rocks arbitrarily grouped by Sobolev as "neivites" are, in effect, gabbroids differing from one another in the ratios of the rock-forming minerals, — plagioclase, hornblende, and pyroxene — and by the degree of idiomorphism in relation to each other.

They all show evidence of a greenstone alteration expressed in albitization of plagioclase and in replacement of dark minerals by a secondary blue-green hornblende. Accordingly, they are characterized by a higher Na_2O content (up to 6%). It is natural, then, that their chemical composition does not correspond to that of the primary magmatic rock, but rather reflects the composition of an altered rock.

Thus, conversion of the chemical analysis data for this rock to numerical characteristics along with the construction of diagrams, and by using them in searching the tables for corresponding rocks — all this led Sobolev to erroneous conclusions on the unique nature of these rocks. All evidence suggests that there are no primary magmatic "neivites"; and that this term, as corresponding to altered gabbro rocks, should be abandoned.

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